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PALEOGEOGRAPHIC AND TECTONIC IMPLICATIONS OF THE LATE PALEOZOIC ALLEGHANIAN OROGEN DEVELOPED FROM ISOTOPIC SEDIMENTARY PROVENANCE PROXIES FROM THE APPALACHIAN FORELAND BASIN

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ABSTRACT OF DISSERTATION

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ABSTRACT OF DISSERTATION

A dissertation submitted in partial fulfillment of the
requirements for the degree of Doctor of Philosophy in the
College of Arts and Sciences
at the University of Kentucky

By
Thomas Patrick Becker

Lexington, Kentucky

Director: Dr. William A. Thomas, Hudnall Professor of Geological Sciences

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The Alleghanian orogeny was a collision between the Gondwanan and Laurentian continents that produced the Pangean supercontinent. Mechanical and kinematic models of collisional orogens are believed to follow a critical taper geometry, where the tectonic imbrication of continental crust begins nearest to the edge of continental plate and advances toward the craton in a break-forward sequence. Studies of shear zones within the Alleghanian collisional orogen, however, suggest that most of the early deformation was translational. Propagation of craton-directed thrusts into the foreland did not occur until the latest Pennsylvanian in the southern Appalachians, and the middle-late Permian in the central Appalachians.

Radiometric sedimentary provenance proxies have been applied to the late Mississippian-early Permian strata within the Appalachian foreland basin to determine the crustal composition and structural evolution of the orogen during the continental collision. U-Pb ages of detrital zircons from the early to middle Pennsylvanian sandstones suggest that most of the detritus within the Appalachian basin was recycled from Mesoproterozoic basement and Paleozoic strata of the Laurentian margin. The presence of Archean and late Paleoproterozoic age detrital zircons is cited as evidence of recycling of the Laurentian syn-rift and passive-margin sandstones. Detrital zircon ages from early-middle Permian-age sandstones of the Dunkard Group do not contain any Archean or Paleoproterozoic detrital-zircon ages, implying a source of sediment with a much more restricted age population, possibly the igneous and metamorphic internides or middle Paleozoic sandstones from the Appalachian basin. The persistance of 360-400 Ma K/Ar ages of detrital white mica suggest that the sediment was supplied from a source that was exhumed during the Devonian Acadian orogeny. Detrital-zircon and detrital-white-mica ages from Pennsylvanian-age sandstones indicate that the late Paleozoic orogen did not incorporate any significant synorogenic juvenile crust. The $^{87}Sr/^{86}Sr$ ratios of middle Pennsylvanian-early Permian lacustrine limestones within the Appalachian basin show a slight enrichment through time, suggesting that labile ^{87}Sr-rich minerals in the Alleghanian hinterland are being exposed. Stable isotopic data from the

lacustrine limestones also corroborates that the Appalachian basin became much more arid through time.

KEYWORDS: Appalachian, Detrital Zircon, Provenance, Pennsylvanian, Permian

________Thomas P. Becker________

________December 16, 2004________

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RULES FOR THE USE OF DISSERTATIONS

DISSERTATION

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This dissertation is dedicated to

My parents, James Arthur and Donna Deanna Becker,

My parents in-law, Bahadur Sarkari, M.D. and Shirin Sarkari, Ph.D.,

and lastly

My incredible wife, Natasha Sarkari Becker, M.D.

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CHAPTER 1. GEOCHRONOLOGICAL FRAMEWORK OF CRUSTAL SEDIMENTARY SOURCES TO THE ALLEGHANIAN OROGENY IN EASTERN NORTH AMERICA

Introduction

The 3000-km long Appalachian mountain chain was produced as part of one of the most significant mountain building events in the geologic record; the assembly of the late Paleozoic Pangean supercontinent (Fig. 1). The Alleghanian orogeny is the collisional event between the eastern margin of present-day North America (Paleozoic Laurentia) and the western margin of Europe, Africa, and South America (Paleozoic Gondwana) during the late Mississippian-late Permian (335-250 Ma). That collision assembled much of Pangea (Fig. 1). The purpose of this chapter is to provide a template for understanding the composition of the orogenic edifice as the provenance for late Paleozoic clastic sediment, the topic of succeeding chapters of this dissertation.

Our conceptual understanding of the late Paleozoic Alleghanian orogeny has been shaped by early studies of the late Paleozoic Appalachian hinterland (e.g., Hess, 1939; Bird and Dewey, 1970; Rodgers, 1970; Hatcher and Williams, 1982) and foreland (e.g., Woodward, 1957; Colton, 1970; Thomas, 1977). Over the past two decades, the tectonic evolution of the late Paleozoic Alleghanian orogeny has been revised substantially, using detailed structural and geochemical studies in the Appalachian hinterland. As a result, the late Paleozoic Alleghanian orogeny is generally accepted to be an oblique collision between Gondwana and Laurentia in the late Paleozoic (e.g. Rodgers, 1987; Gates et al., 1988; Shelley and Bossiere, 2000; Hatcher, 2002). However, an integrative approach that reconciles the kinematic and tectonic history of the Alleghanian hinterland to the sedimentary record preserved in the Appalachian basin does not exist. The goal of this dissertation is to provide a framework to evaluate the sedimentary record in the

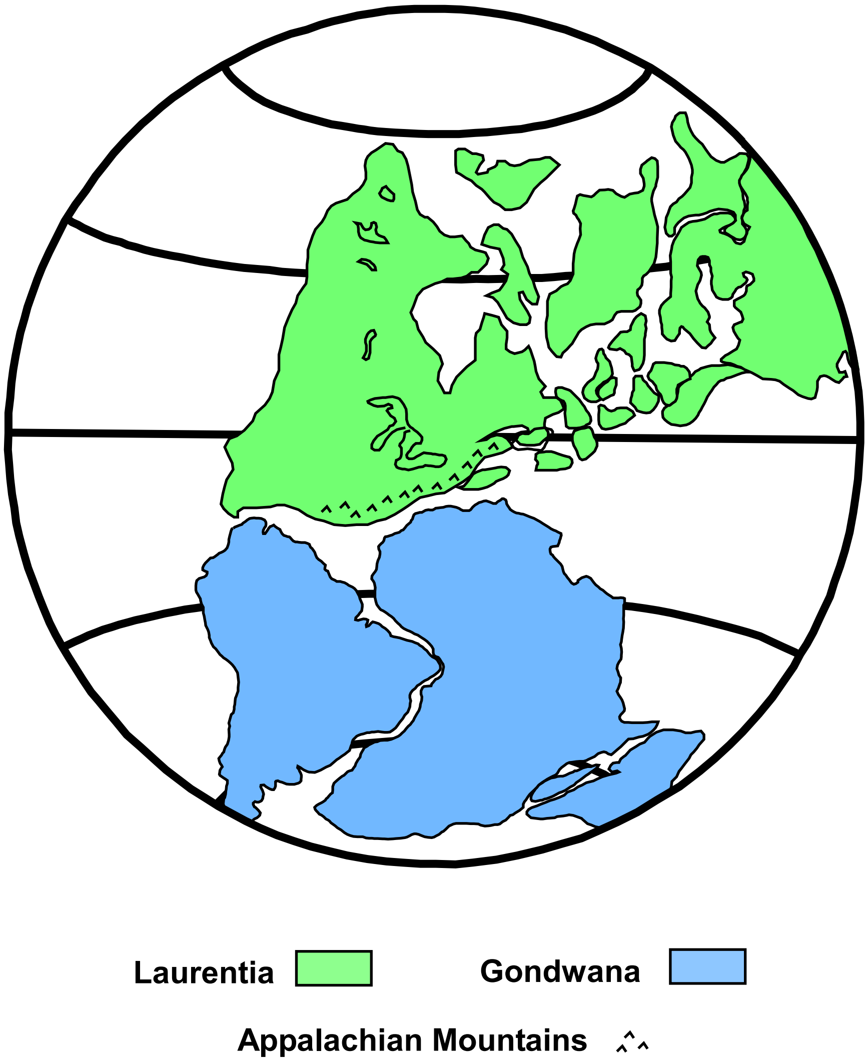

Figure 1.1. Reconstruction of the Pangean supercontinent at about 300 Ma (late Pennsylvanian) based on paleomagnetic data (modified from Van der Voo and Torsvik, 2001)

Appalachian basin in the context of our evolving understanding of the late Paleozoic Alleghanian orogenic hinterland.

Overview of Pangea

The concept of late Paleozoic continental amalgamation into a large supercontinent was created along with the plate-tectonic paradigm by Alfred Wegener in 1915 (Wegener, 1970), to explain the seeming likeness between the North American, South American, and African coasts, and the distribution of late Paleozoic and early Mesozoic flora. The Pangean concept became a central tenet as plate tectonics matured from hypothesis to theory to paradigm in the 1960s. The Pangea hypothesis was supported strongly by the simple geometry of Pangean continental fragmentation along relatively fixed Euler poles throughout the Mesozoic (Wilson, 1966; Francheteau and LePinchon, 1972). In one of the early defining papers of plate tectonics, Bullard et al. (1965) matched the North American, South American, and African continental margins of the -900 m Atlantic contour, and produced a fit that has good agreement with modern paleomagnetic data (Van der Voo, 1990, 1993). This provided the conceptual groundwork for evaluating the Appalachians in the context of a plate-tectonic collision between eastern Laurentia (North America) and western Gondwana (western South America and Africa).

Early (and influential) papers regarded the Appalachians as the result of a plate tectonic collision implied orthogonal convergence between Laurentia and Gondwana in the late Paleozoic to produce the supercontinent Pangea (Wilson, 1966; Bird and Dewey, 1970; Hatcher, 1987). This interpretation was supported by the observed bilateral symmetry of structures on Laurentia and Gondwana around the core of the mountain belt, and the vergence of thrust belts away from the collisional front. In addition, the sedimentary record in the Appalachian basin shows a rapid transition in the Appalachian basin from a shallow-marine embayment in Mississippian time to an eastward-thickening clastic wedge in the Pennsylvanian (Ferm, 1974; Thomas, 1977). The early conceptual model of orthogonal convergence represented an attempt to apply the plate-tectonic model in its simplest form to the formation of the Appalachian Mountains.

Orthogonal collision between plates requires the destruction of oceanic lithosphere along the margin of one of the continents involved in the collision. Typically, this results in the formation of an accretionary prism and volcanic arc terranes along the border of the continental hanging wall in the subduction zone that brings the continental masses together. Neither arc terranes nor an accretionary prism have been convincingly identified in the late Paleozoic Appalachians (Samson et al., 1995; Coler et al., 1997). In addition, many lithologic features that generally are associated with continental collision, such as obducted ophiolites and exhumed ultra-high pressure rocks, are not preserved in the geologic record of the late Paleozoic Appalachian collision.

Presently, most Appalachian (and Hercynian) geologists conclude that the collision between Gondwana and Laurentia was highly oblique (e.g., Gates et al., 1986, 1988; Gromet, 1989; Hatcher, 1989, 2002; Shelley and Bossiere, 2000; Valentino and Gates, 2001; Vai, 2003). This is attributed to the recognition of numerous dextral shear zones that parallel the continental margin within the Appalachian crystalline hinterland (Fig. 2). These shear zones comprise a broad zone of dextral displacement that extends from Alabama to Newfoundland (Gates et al., 1988; Edleman et al., 1987; Reck and

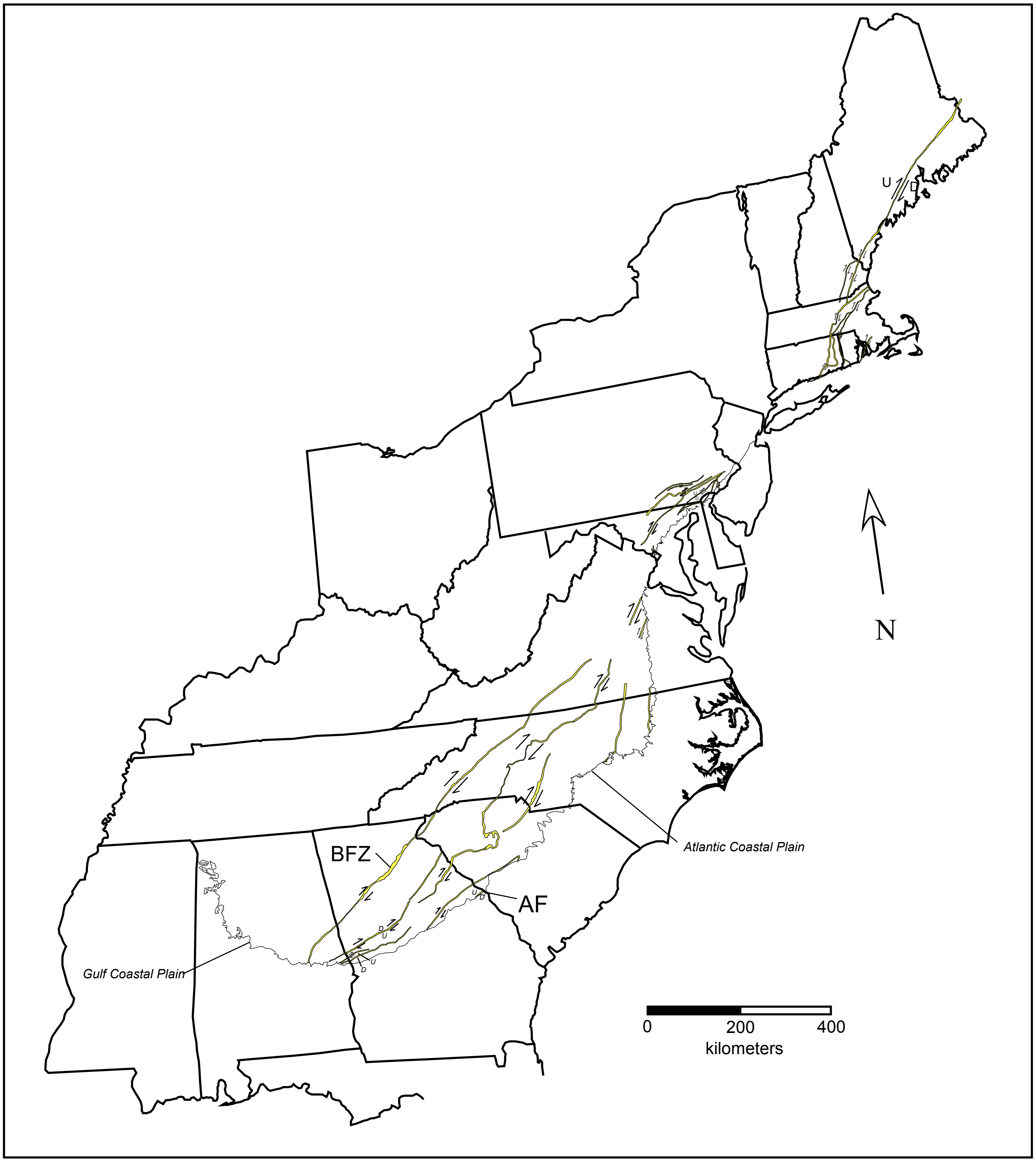

Figure 1.2. Distribution of dextral shear zones in the Appalachian hinterland. Compiled from Gates et al. (1988), Steltenpohl et al. (1992), Valentino and Gates (2001), Goldstein and Hepburn (1999). BFZ refers to the Brevard fault zone, and AF is the Augusta fault.

Mosher, 1988; West and Lux, 1993; Steltenpohl et al., 1992; Hibbard et al., 1998; Valentino and Gates, 2001). Estimates of the total displacement along individual Appalachian dextral shear zones range from tens (Gates et al., 1988) to hundreds (Druhan et al., 1988; Gates et al., 1988) of kilometers. Attempts to constrain the timing of movement by radiometric means ($^{40}Ar/^{39}Ar$ and U-Pb geochronology) along these dextral shear zones yield a range of ages from 335 to 270 Ma (Zartman and Hermes, 1987; Stockey and Sutter, 1991; Steltenpohl et al., 1992; West and Lux, 1993; Wright and Dennis, 1995; Wortman et al., 1998; Krol et al., 1999). Because of the pervasive nature and multiple episodes of dextral displacement (and lack of evidence of a subduction margin), these shear zones likely characterize the Alleghanian orogeny, and are not merely a secondary consequence of collision.

The late Paleozoic tectonic history of Europe and northern Africa (tectonic complement to Laurentia) reflects a similar construction of Pangea. Prior to the formation of Pangea, a proto-Pangea (or Pangea B, Fig. 3) formed as northwestern Gondwana collided obliquely with the southeastern margin of Laurentia along dextral transcurrent shear zones in central and western Gondwana during the late Pennsylvanian-early Permian (Vai, 2003). This Pangean configuration apparently was unstable and was broken by further dextral shear, possibly causing the formation of several early Permian rift basins in northern Europe (Oslo-Polish rift basin) and the eastern Meditteranean (Oman/Iraq/Levantine/Sicily basin) (Fig. 3) (Ziegler, 1990; Angiolini et al., 2003; Maystrenko et al., 2003; Stovba et al., 2003; Vai, 2003). Several intramontane pull-apart basins also formed in the late Pennsylvanian-early Permian within central Europe (Vai, 2003), eastern Greenland (Coffield, 1993), and northeastern North America (Fig. 3)

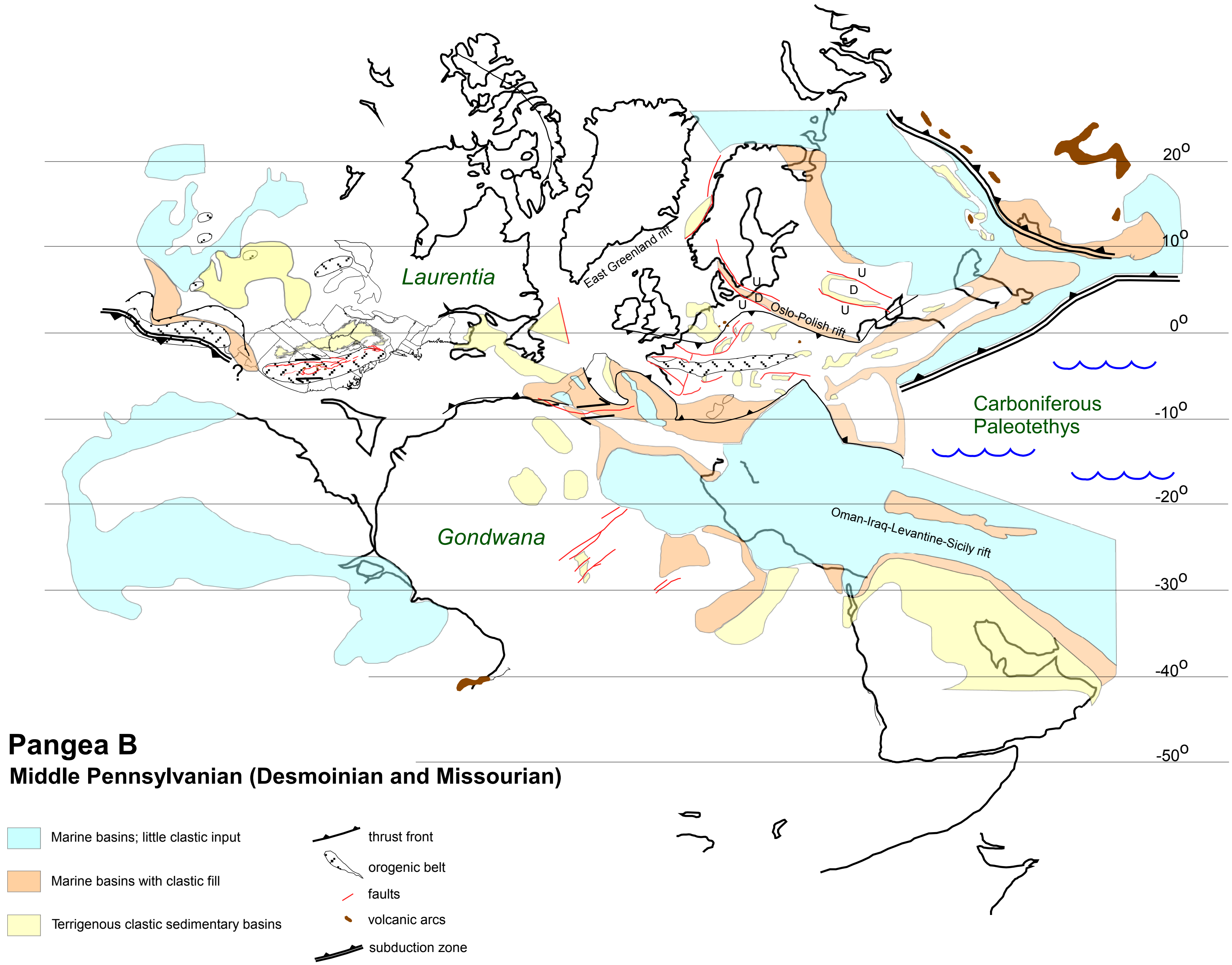

Figure 1.3. Map of the middle Pennsylvanian Pangea B supercontinental configuration, modified from Vai (2003).

(Bradley, 1982; Thomas and Schenk, 1988). Preliminary data from seismic reflection profiles also hint at the presence of an early Permian rift basin in northwestern Africa (Vai, 2003). Alkaline volcanic rocks associated with a rift system were also emplaced along a trend from the British Isles, to the Oslo graben, and into Poland from 300 to 275 Ma (Neuman et al., 1985; Ziegler, 1988; Olaussen et al., 1994). Additional mafic dikes were emplaced from 246 to 238 Ma during the terminal stages of the Oslo rift (Torsvik et al., 1998). This oblique rifting event is interpreted to have separated Laurentia from Gondwana and allowed rotation of the continents until they recombined in the late Permian to form the much more stable Pangea A assembly (Fig. 4) (Shelley and Bossiere, 2000; Vai, 2003). In the Pangea A assembly of Vai (2003), northwestern Africa rotated to collide with northeastern North America. By the late Permian, Laurussia had also accreted to Baltica along the Uralian orogenic front, establishing the most complete assembly of Pangea. Paleontological evidence supports a stable Permian Pangean assembly, as well. Parts of the Laurentian provincial fauna did not appear in Gondwana until the latest Permian/early Triassic (Vai, 2003), adding evidence that the Pangean supercontinent was not fully assembled until the very end of the Paleozoic.

Our understanding of the Pangean tectonic assembly is continually evolving, and models of the Alleghanian orogeny have to accomodate this evolution. One approach that can be used to provide constraints on the tectonic history of the Alleghanian orogeny is the use of radiometric indicators of sedimentary provenance on clastic deposits in the Appalachian and Black Warrior basins. The detrital zircons can be used to identify the provenance of sediment that was shed into the foreland basins. The results can be examined in the context of the kinematic history between Gondwana and Laurentia in the

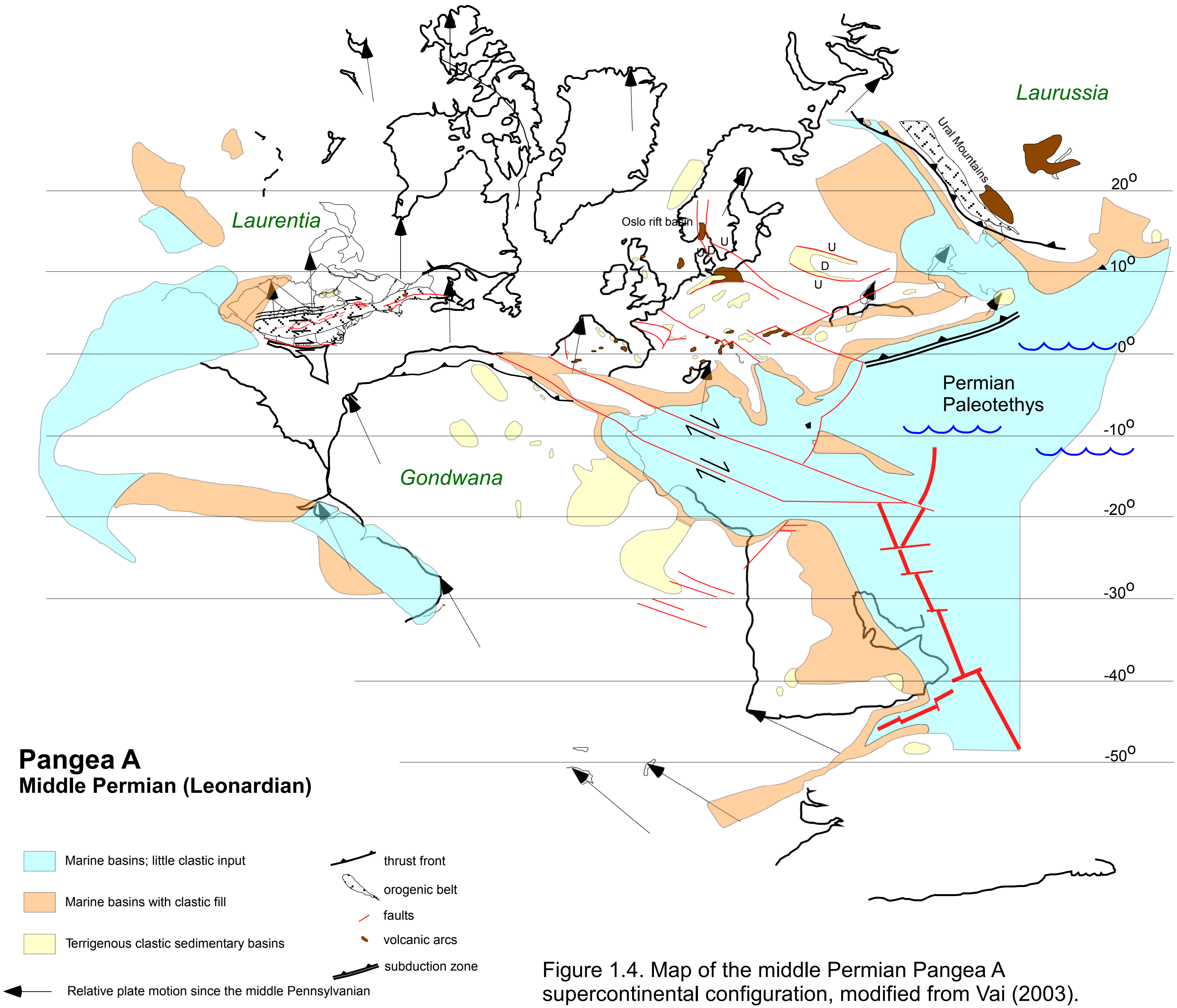

Figure 1.4. Map of the middle Permian Pangea A supercontinental configuration, modified from Vai (2003).

late Paleozoic. The following discussion provides a tectonic context for the interpretation of detrital-zircon ages from sedimentary deposits in the Appalachian basin.

Possible sediment sources for detrital zircons in Appalachian (Alleghanian) Pennsylvanian sandstones

U-Pb dating of detrital zircons has revolutionized studies of sedimentary provenance. In conventional sedimentary petrography, provenance was assessed by determining the relative proportions of constituents of sandstone framework grains (Fig. 5) (Dickinson and Suzcek, 1979; Dickinson et al., 1983). This approach yields reproducible data, but the determination of provenance is not uniquely constrained. In addition, chemical alteration during production, transport, and diagenesis of sediment may significantly alter its composition so virtually no unique lithologic attributes of the provenance are retained in the sediment (e.g., Johnsson, 1993). Detrital zircons are chemically and mechanically resistant to weathering, and are thought to be transported along with the most chemically resistant mineralogic phases, such as quartz and tourmaline (Morton and Hallsworth, 1994). U-Pb ages of detrital zircons provide a very reliable means of determining the crystallization age of sedimentary sources, which can be tied to source regions, without the loss of critical information because of weathering, transport, and diagenetic processes.

Zircons are produced petrogenetically in igneous and granulite-grade metamorphic rocks, which are commonly associated with tectonism. In order to assess provenance from detrital-zircon ages, it is necessary to have an understanding of the spatial and temporal distribution of various crustal terranes created in earlier tectonic events and their likely involvement in subsequent tectonism. The following section is

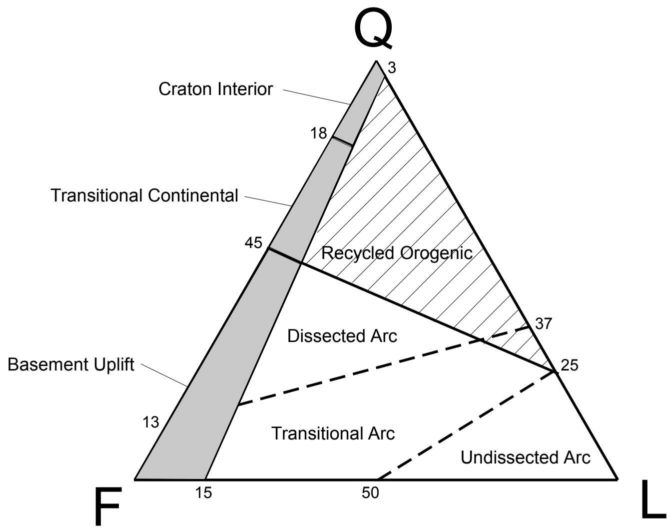

Figure 1.5. Empirically derived ternary diagram of Dickinson et al. (1983) showing the relationship between framework grain composition and tectonic setting. Q: all quartz (including chert), F: Feldspar (alkali and calcium), L: all lithic fragments.

organized by ages of zircon-producing tectonic events and their geographic distribution within Laurentia and peri-Gondwanan terranes. Two types of source rocks are being considered; those that crystallized during a particular tectonic episode (primary sources), and those that may have experienced multiple episodes of crustal recycling (recycled sedimentary sources).

Primary Crystalline Sources

1600-1950 Ma , >2500 Ma

The oldest continental crust in North America is restricted to a few discrete areas within North America (Paleozoic Laurentia), the most ancient of which is found in northwestern Canada and in western Wyoming, comprising the Wyoming Province (2600-2900 Ma) (Fig. 6). The Wyoming and the vast Superior Province (2600-2800 Ma) of eastern and central Canada were conjoined during the 1800-1900 Trans-Hudson and Penokean orogenies (Fig. 6) (Hoffman, 1989; Van Schmus et al., 1993). The Superior-Wyoming continental mass was a nucleation site for the growth of the North American continent during subsequent early to middle Proterozoic Yavapai-Mazatzal and Central Plains plate collisions along its southern margin from 1600 to 1800 Ma (Fig. 6) (Hoffman, 1989).

1300-1500 Ma

The Granite-Rhyolite Province (Fig. 6) is a suite of granitic and rhyolitic igneous rocks in the central and southern United States (Anderson, 1983; Hoffman, 1989; Bickford et al., 2000). They are interpreted to be anorogenic in origin, but they are widespread (Fig. 6). Most of the Granite-Rhyolite province is covered by early-middle Paleozoic sedimentary rocks (Rankin et al., 1989), precluding this region from directly

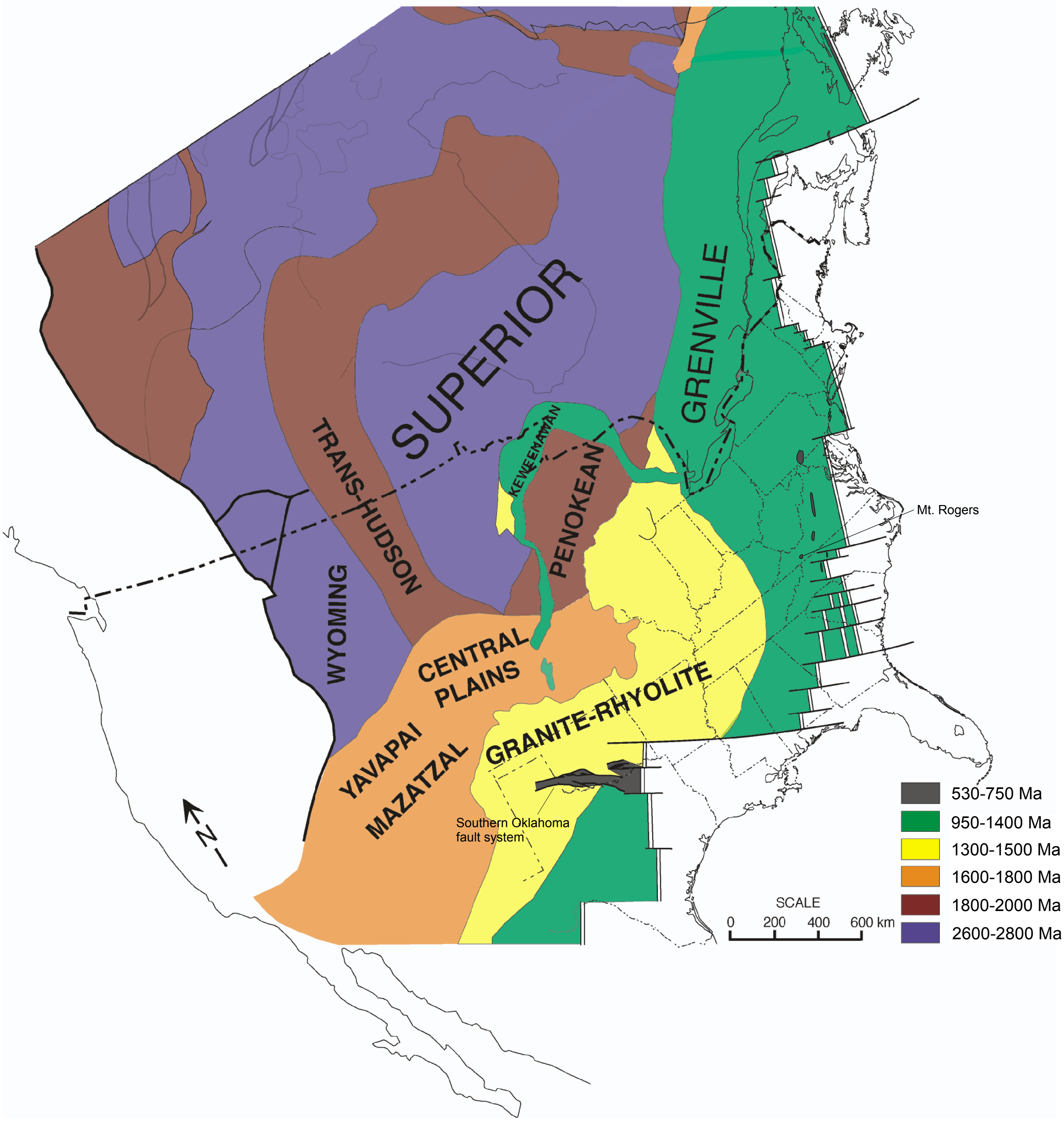

Figure 1.6. Map of the eastern half of North America showing the age and distribution of terranes that comprised Laurentia in the late Proterozoic/early Paleozoic. Data compiled from Bickford et al. (1986); Hoffman (1989); Van Schmus et al. (1993).

supplying sediment to the Appalachian basin from the early Paleozoic through the late Paleozoic.

The regions of Laurentian crust with crystallization ages >1300 Ma probably did not supply sediment directly to the Appalachian foreland basins (Appalachian and Black Warrior basins). Progressive onlap of the early Paleozoic passive margin across the Canadian shield likely prevented the crystalline shield from supplying sediment directly to the Appalachian basin in the late Paleozoic (Rankin et al., 1989). This assumption is supported by detrital-zircon-age populations of the lower and middle Paleozoic sedimentary rocks from Virginia and New York (McLennan et al., 2001; Eriksson et al., 2004). In contrast, much of the sediment that filled the Eastern Interior (Illinois) basin was likely derived from the shield (Potter and Pryor, 1961; Collinson et al., 1988).

950-1250 Ma (Grenville)

Metamorphic and plutonic rocks of the Grenville orogenic belt define assembly of the Rodinia supercontinent at ~1250 to 950 Ma (e.g., Dalziel, 1997). The metamorphic terranes encompass enclosures of older rocks ranging to 1700 Ma; however, these are more common in the north (Labrador) than farther south (Fig. 6) (Cawood and Nemchin, 2001). The subsequent breakup of the Rodinian supercontinent and opening of Iapetus Ocean left a band of rocks of the Grenville province along the eastern rim of Laurentia (Fig. 6), where the Grenville rocks became the continental “basement” to the Paleozoic Appalachian orogenies. Grenville rocks are incorporated in Appalachian external and internal basement massifs, and the precursors of the present eroded massifs may have been primary sources of detritus supplied directly to the Alleghanian foreland basin. In addition, many of the paragneisses that comprise the Appalachian crystalline hinterland have detrital-zircon populations that are wholly Grenville in age (Bream, 2002; Bream et al., 2004).

550-750 Ma

Diachronous rifting and opening of the Iapetus Ocean is documented by igneous rocks that range from ~620 Ma in Newfoundland to 530 Ma along the Southern Oklahoma fault system (Fig. 6) (Hogan and Gilbert, 1998; Thomas et al., 2000; Cawood and Nemchin, 2001). The geographically most extensive rifting (along the Blue Ridge rift, Fig. 6) is marked by synrift igneous rocks of 572±5 to 564±9 Ma (Aleinikoff et al., 1995; Walsh and Aleinikoff, 1999).

An earlier phase of rifting along the southern Appalachian Laurentian margin has also been proposed by Aleinikoff et al. (1995) to explain the distribution of ~750 Ma granites in western North Carolina and Virginia (Fig. 6). Lukert and Banks (1984) reported a zircon U-Pb age of 732 +/- 5 Ma for the Robertson River pluton, north of Charlottesville, in central Virginia. The Robertson River pluton may be related to ~750 Ma volcanism at Mount Rogers (Aleinikoff et al., 1995), the ~740 Ma Crossnore Complex (Su et al., 1994), and the 734 +/- 26 Ma Bakersville granite/gabbro intrusive suite in North Carolina (Goldberg et al., 1986).

Field relationships show that the ~750 Ma igneous rocks are of limited exposure and extent and are unconformably overlain by lower Paleozoic strata, so that there is little reason to believe that these Neoproterozoic volcanic rocks were a substantial source of detritus into the Appalachian basin. In contrast, the Iapetan synrift rocks (530-620 Ma) are exposed in basement-cored uplifts from Newfoundland to North Carolina.

Neoproterozoic detrital-zircon ages (600-800 Ma) are most commonly attributed to Gondwana (see below), but some of these detrital zircons could be of Laurentian origin.

440-490 Ma

Plutons in the Appalachian Piedmont (~440-490 Ma) are interpreted to be the eroded roots of Ordovician volcanic systems that were contemporaneous with the Taconic orogeny in New England (Karabinos et al., 1998; Sevigny and Hanson, 1993; Tucker and Robinson, 1990), the central Appalachians (Shaw and Wasserburg, 1984; Sinha et al., 1997; Aleinikoff et al., 2002), and the southern Appalachians (Coler et al., 2000; McClellan and Miller, 2000; Miller et al., 2000; Hibbard, 2000). Related Taconic metamorphism temporally overlaps magmatism in western New England (Sutter et al., 1985; Ratcliffe et al., 1998), the central Appalachians (Bosbyshell et al., 1998), and the southern Appalachians (Miller et al., 2000).

350-420 Ma

Plutons contemporaneous with the Acadian orogeny (~350-420 Ma) are distributed along the Appalachian Piedmont (Osberg et al., 1989; Miller et al., 2000), although they are concentrated in the northern Appalachians (Osberg et al., 1989; Eusden et al., 2000). Regional metamorphism contemporaneous with the magmatism has been documented in New England (Hames et al., 1991; Eusden et al., 2000) and the southern Appalachians (Osberg et al., 1989).

330-280 Ma

Several granitic plutons within the Eastern Blue Ridge, Piedmont, and accreted Gondwanan terranes of the southern Appalachians were emplaced during the Alleghanian orogeny from 335 to 300 Ma (Fig. 7) (Wright et al., 1975; Dallmeyer et al., 1986; Horton et al., 1987; Dennis and Wright, 1995; Heatherington and Mueller, 1997; Heatherington, 1998; Schneider and Samson, 2001; Samson, 2001; Miller et al., in review). The central Appalachians do not have any significant igneous intrusive bodies of Alleghanian age. It is possible that they do exist, but are covered by the Mesozoic-present Atlantic Coastal Plain strata. Several late Paleozoic granites within southeastern New England include the Sebago batholith in eastern Maine (Tomascak et al., 1996), the Narragansett Pier Granite (Zartman and Hermes, 1987), the Potter Hill granite of eastern Connecticut (Whitehead and Gromet, 1997), and the Pinewood Adamellite of eastern Connecticut (Fig. 8) (Sevigny and Hanson, 1993). Several other granitic plutons (Lyman, Saco, Biddeford, Mason, Hollis) are possibly of late Paleozoic age, but only ^{87}Rb-^{87}Sr ages presently are available (Gaudette et al., 1975; Hayward and Gaudette, 1984).

Many of these late Paleozoic plutons have been difficult to identify because of substantial inheritance of older xenocrystic zircon cores, as a result of zirconium saturation in their melts (Zartman and Hermes, 1987; Whitehead and Gromet, 1997; Heatherington and Mueller, 1997; Miller et al., in review). Inheritance of xenocrystic zircons may be attributed to the source of melt to the late Paleozoic plutons, which is hypothesized to be remelted country rock (Samson et al., 1995; Coler et al., 1997). In detrital-zircon-age populations, contributions from Alleghanian-age plutons might be difficult to detect because the cores of the zircons are typically analyzed.

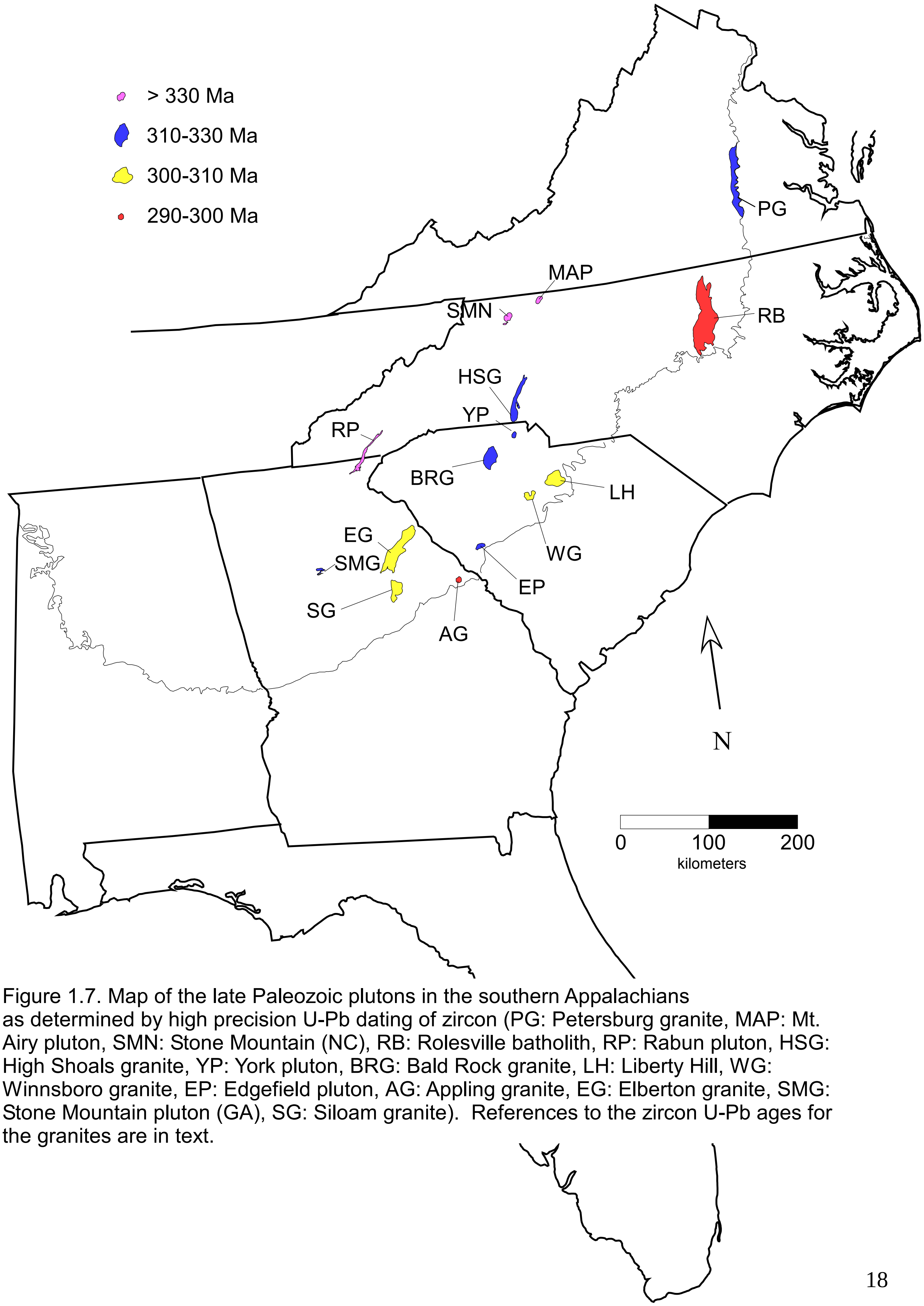

Figure 1.7. Map of the late Paleozoic plutons in the southern Appalachians as determined by high precision U-Pb dating of zircon (PG: Petersburg granite, MAP: Mt. Airy pluton, SMN: Stone Mountain (NC), RB: Rolesville batholith, RP: Rabun pluton, HSG: High Shoals granite, YP: York pluton, BRG: Bald Rock granite, LH: Liberty Hill, WG: Winnsboro granite, EP: Edgefield pluton, AG: Appling granite, EG: Elberton granite, SMG: Stone Mountain pluton (GA), SG: Siloam granite). References to the zircon U-Pb ages for the granites are in text.

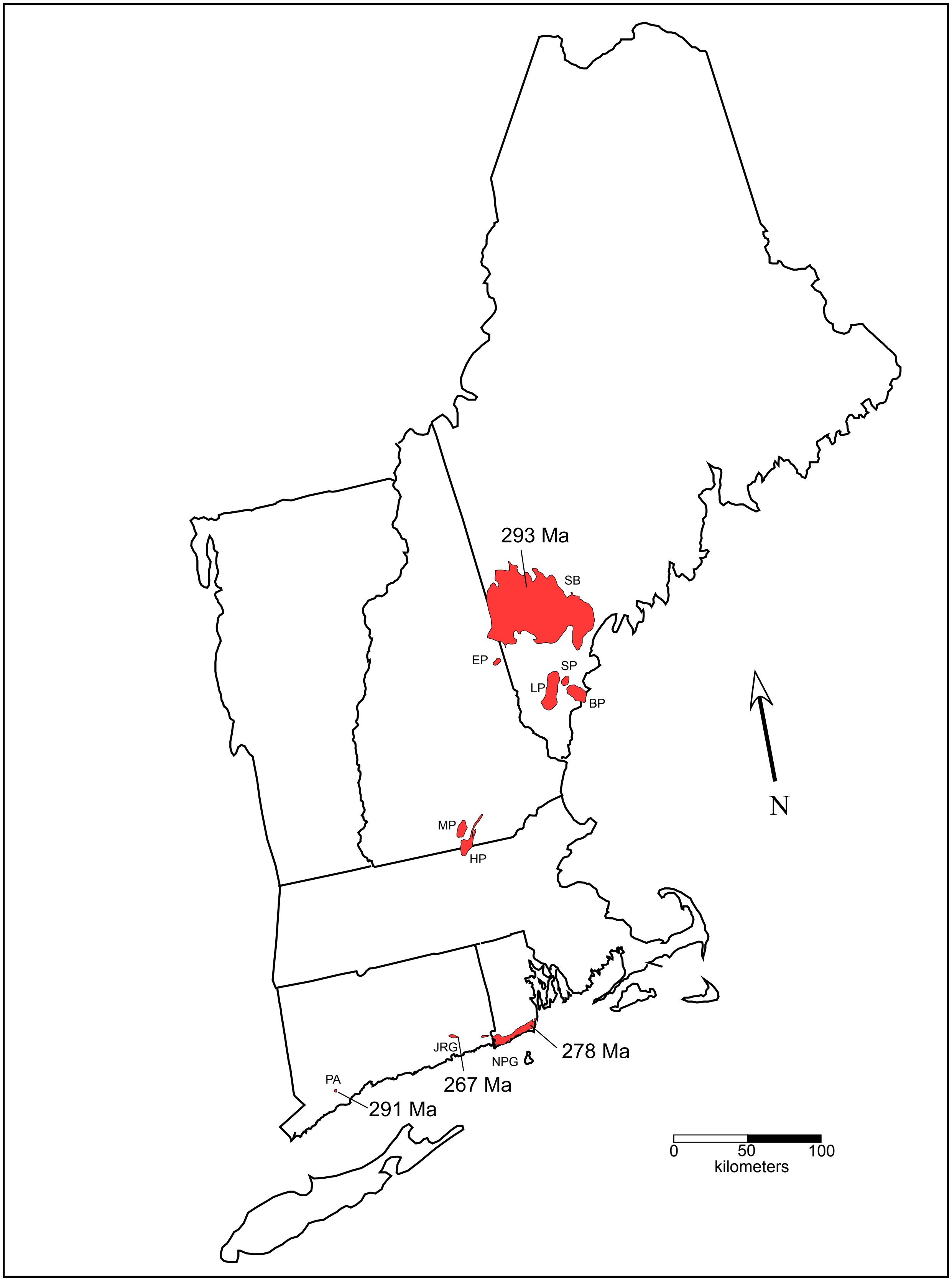

Figure 1.8. Late Paleozoic granitic intrusions in New England. The plutons are: SB-Sebago batholith, EP-Effingham pluton, SP-Saco pluton,LP-Lyman pluton, BP-Biddeford pluton, MP-Mason pluton, HP-Hollis pluton, NPG-Narragansett Pier granite, JRG-Joshua Rock granite gneiss, PA-Pinewood Adamellite. Ages associated with some plutons represent high-precision U-Pb dates that record timing of emplacement. Other pluton ages are based on the ^{87}Rb-^{87}Sr method. References to ages are in text.

520-660 Ma; 1950-2200 Ma (Gondwanan)

Several crustal terranes (e.g. Wiggins, Suwanee, Carolina, Smith River; Fig. 9) that border the Atlantic coast in the southern Appalachians have been recognized as being of Gondwanan affinity on the basis of faunal correlation (Secor et al., 1983), magmatic history (Samson, 1995), and detrital-zircon ages (Mueller et al., 1994; Ingle-Jenkins et al., 1998; Eriksson et al., 2004). Because these terranes have Gondwanan crustal characteristics, it is difficult to determine whether they represent remnants of Alleghanian or earlier Paleozoic accretion. Hibbard (2000) and Hibbard et al. (2002) highlighted several lines of geologic evidence that support a Silurian-Ordovician collision for the Carolina terrane, which is in agreement with previous interpretations (Glover et al., 1983; Horton et al., 1989). Detrital-zircon ages from the Aaron and Uwharrie Formations within the Carolina terrane have a single mode at 510-620 Ma (Ingle-Jenkins et al., 1998; Eriksson et al., 2004). On the basis of chemical age dates of monazites, Hibbard et al. (2003) interpreted the Smith River terrane to be Gondwanan and to have been accreted during the late Ordovician, however, no zircon ages are available from the Smith River allochthon. The Wiggins and Suwanee terranes are interpreted to be remnants of Gondwana that were accreted during the Alleghanian orogeny and that remained attached to North America following the opening of the Atlantic Ocean (Horton et al., 1989). Mueller et al. (1994) analyzed detrital zircons from Silurian-Devonian strata from the Suwannee terrane and found a bimodal population of detrital-zircon ages at 515-637 Ma and 1967-2282 Ma, consistent with the interpretation of a Gondwanan origin for the Suwannee terrane.

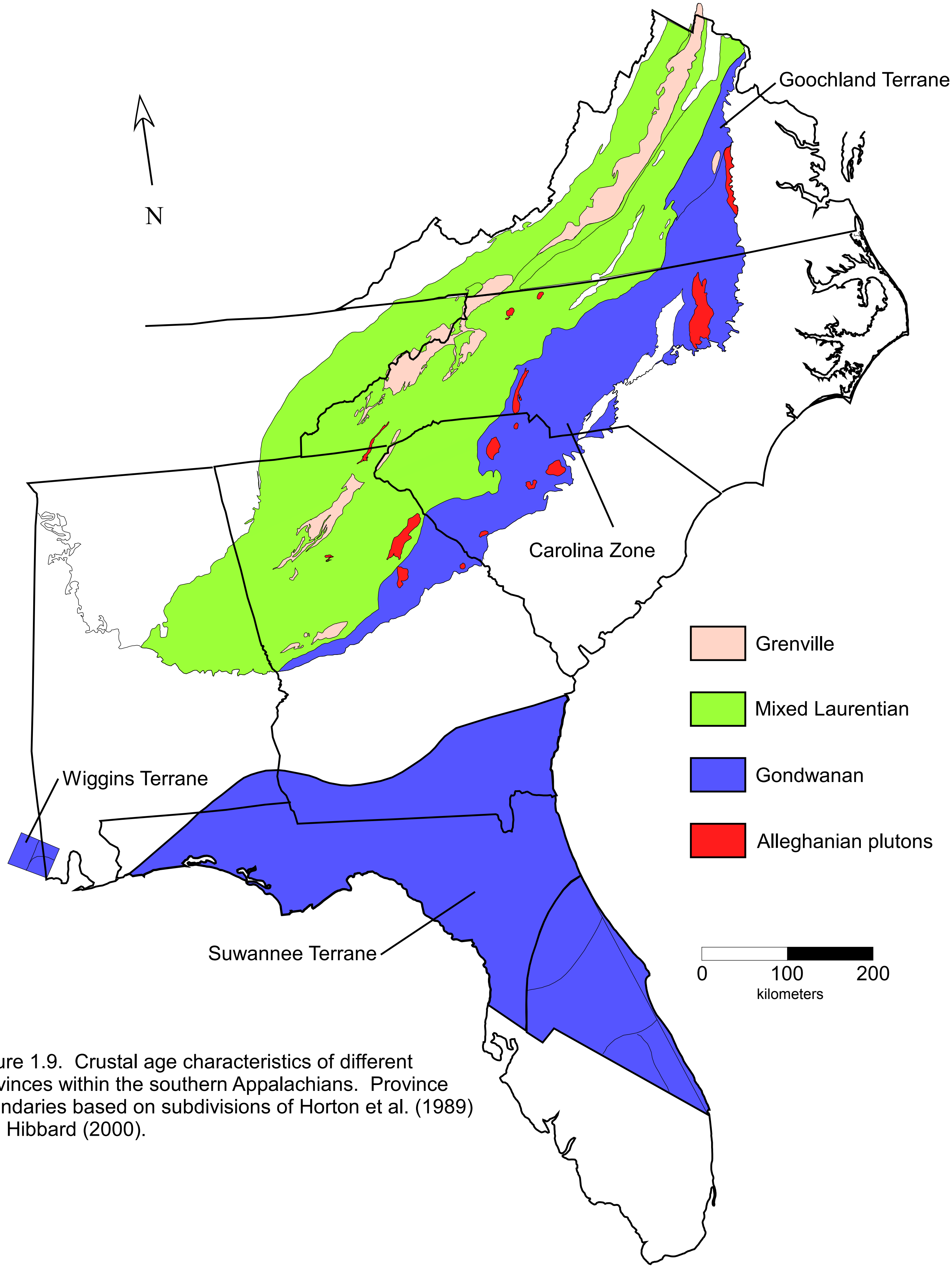

Figure 1.9. Crustal age characteristics of different provinces within the southern Appalachians. Province boundaries based on subdivisions of Horton et al. (1989) and Hibbard (2000).

The Goochland terrane, in eastern Virginia (Fig. 9), hosts plutons of Neoproterozoic age (654 to 588 Ma) that are intruded into a Grenville-age basement (Owens and Tucker, 2003). Although Owens and Tucker (2003) interpret the ages to represent late Proterozoic rifting of a continental block of Laurentian origin, the juxtaposition of the Goochland terrane to the east of the Ordovician Milton and Chopwamsic arc terranes (Coler et al., 2000), and the timing of magmatism coincident with that in Gondwana support a Gondwanan origin. Crust of Gondwanan affinity is also likely present beneath the Atlantic Coastal Plain. Recent drilling of an Eocene bolide impact structure in the Chesapeake Bay revealed that the crystalline basement is composed of a peraluminous, weakly metamorphosed 614 +/- 9 Ma granite (Horton et al., 2001). The Neoproterozoic age overlaps with the Pan-African/Brasiliano orogenic event in Gondwana.

In eastern New England, a combination of geological and geochronological data indicates that the peri-Gondwanan Avalon terrane was accreted during the Paleozoic (Fig. 10) (Robinson and Hall, 1980; O'Hara and Gromet, 1985; Dorais et al., 2001). A number of geochronological studies confirm that the Avalon terrane is composed of crust that is of a vintage more typical of Gondwana (580-660 Ma) (reviewed in Skehan and Rast, 1990). Detailed study of the Avalon terrane by O'Hara and Gromet (1985) resulted in further subdivision of the Avalon terrane into the Esmond-Dedham and Hope Valley terranes. The Hope Valley terrane underplated the eastern margin of Laurentia (Wintsch and Aleinikoff, 1987; Getty and Gromet, 1992a,b), and is separated from the western Esmond-Dedham terrane by the Hope Valley shear zone (O'Hara and Gromet, 1985). The Hope Valley terrane is interpreted to have collided with Laurentia during the

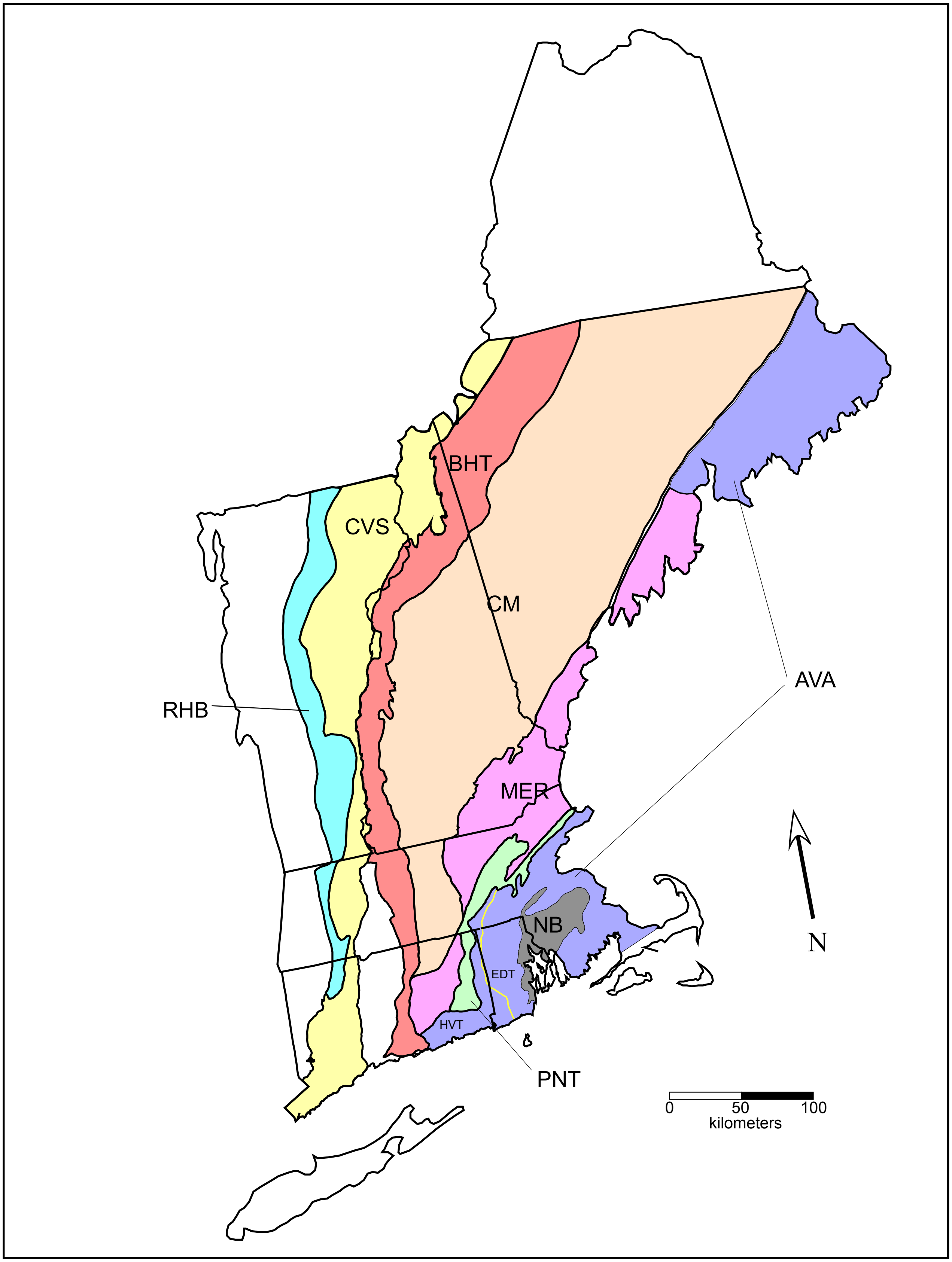

Figure 1.10. Lithotectonic terranes of New England after Wintsch et al. (2003). The terranes are : RHB-Rowe-Hawley belt, CVS-Connecticut Valley synclinorium, BHT-Bronson Hill terrane, CM- Central Maine terrane, AVA- Avalon terrane, MER-Merrimack terrane, PNT-Putnam-Nashoba terrane, NB-Narragansett basin. HVT is the Hope Valley Terrane and EDT is the Esmond-Dedham terrane of O'Hara and Gromet (1985).

Devonian or Ordovician (Hermes and Zartman, 1985; Dorais et al., 2001). Recent petrological and isotopic characterization of the Massabesic anticlinorium in southern New Hampshire (Dorais et al., 2001) suggests that the Avalon terrane is one single microcontinent accreted during the Devonian Acadian orogeny, and that the Hope Valley terrane represents its leading edge. The Hope Valley terrane is characterized by granite gneisses and metasedimentary rocks (e.g. O'Hara and Gromet, 1985; Hermes and Zartman, 1992), that are interpreted to be metamorphosed shelf/slope facies of the Avalon terrane (Dorais et al., 2001). The thicker and lithologically diverse Esmond-Dedham terrane is interpreted to be of continental affinity that was accreted along with the Hope Valley terrane during the Devonian, and was thrust into its present position along the Honey Hill fault during the Alleghanian orogeny (O'Hara and Gromet, 1985; Dorais et al., 2001).

Several of the exotic terranes that were accreted to the Laurentian margin during the Paleozoic orogenies (Secor et al., 1986; Dallmeyer, 1988; Horton et al., 1989; Getty and Gromet, 1992a,b) have crustal components with crystallization ages that contrast with the geologic history of Laurentia (e.g., Skehan and Rast, 1990; Mueller et al., 1994; Samson et al., 2001). Timing of accretion of the various terranes is poorly constrained, but estimates range from Taconic (possible Ordovician accretion of the Carolina terrane, Hibbard, 2000) to Alleghanian (Pennsylvanian-Permian accretion of the African Suwannee and Wiggins terranes, Thomas, 1989). The terranes include zircons of distinctive ages, such as 2200 Ma Gondwana basement to Pennsylvanian-Permian granites in the Suwannee terrane (Heatherington et al., 1999) and 550-630 Ma volcanic rocks in the Carolina terrane (Wortman et al., 2000; Hibbard et al., 2002; Eriksson et al., 2004).

Recycled sedimentary sources (mixed Laurentian)

Synrift

Thick accumulations of late Proterozoic-early Paleozoic sediment reflect Iapetan rifting along the eastern Laurentian margin (summary in Thomas, 1991). Potential sources of synrift sediment include contemporaneous synrift volcanic and plutonic rocks, Grenville-age basement on the rift shoulders, and older rocks of the distant Laurentian shield. Analyses of detrital zircons from Appalachian synrift sediment presently are available from Newfoundland, and from the central and southern Appalachians.

Cawood and Nemchin (2001) analyzed three samples from synrift deposits in Newfoundland. The depositional age of the late Precambrian synrift strata is roughly synchronous with the local volcanic/plutonic rocks documented by detrital-zircon ages of 572-628 Ma. A detrital-zircon age of 760±40 Ma contrasts with the ages of nearby synrift igneous rocks but is coeval with the older phase of volcanic rocks along the Blue Ridge rift at Mount Rogers (Fig. 6). The synrift samples are dominated by Grenville (999-1165 Ma) and older zircons with age clusters at 1230-1360, 1740-1870, and 2600-2890 Ma (Cawood and Nemchin, 2001). The older detrital zircons require sediment transport from the distant Laurentian shield to the rift at the Iapetan margin (Fig. 6) (Cawood and Nemchin, 2001). Representation of the older phase of the southern Blue Ridge rift may represent axial sediment transport (Cawood and Nemchin, 2001), or reworking of a local ash deposit. Two of the Newfoundland synrift sandstone samples are similar in containing zircons of a wide range of older ages; however, one sample lacks zircons older than Grenville.

The Neoproterozoic Unicoi Formation in southwestern Virginia is also interpreted to be a synrift sedimentary deposit. The detrital-zircon-age population is dominated by a 950-1250 Ma (Grenville) mode (Eriksson et al., 2004). The remainder of the detrital zircons have ages that range from 1326 to 1427 Ma, 1707 to 1734 Ma, and a single grain of 3296 Ma (Eriksson et al., 2004).

In the southern Appalachians, the Ocoee Supergroup has been interpreted to represent a Late Proterozoic synrift sedimentary deposit (e.g. Rast and Kohles, 1986). Bream (2002) analyzed detrital zircons from the Ocoee, showing that Grenville-age zircons (950-1250 Ma) dominate the population, which does not include any detrital-zircon ages that are contemporaneous with Mount Rogers (~750 Ma) or Iapetan rifting (530-620 Ma).

Local heterogeneity in detrital-zircon-age population indicates substantial variations in the location of drainage systems and dispersal points into the rift system through time. Furthermore, synrift sedimentary deposits are sporadically distributed along the rifted margin, and are thin to lacking along some segments (Thomas, 1991).

Rift-Drift transition

A basal sandstone records a diachronous transition from synrift strata to passive-margin deposits along the Appalachian Iapetus margin (e.g., Thomas, 1991; Cawood et al., 2001); initial transgression of the basal passive-margin sandstone ranges in age from earliest Cambrian (543 Ma) to early Late Cambrian, marking the end of active rifting.

The passive-margin shelf is dominated by carbonate rocks, but some quartzose sandstones are interbedded in the carbonates (e.g., Rankin et al., 1989). The passive-margin sandstones might represent reworking of the basal sandstone or sedimentary transport of detritus onto the shelf from the center of the craton.

The basal (rift-drift transition) sandstone (Bradore Formation) in Newfoundland contains detrital zircons exclusively of Grenville age (Cawood and Nemchin, 2001). The basal sandstone (Poughquag Quartzite) in New York is dominated by Grenville-age zircons, but the sample also contains zircons with dates of 1300-1540 Ma, suggesting derivation from either older enclaves within the Grenville orogen or the mid-continent Granite-Rhyolite province (McLennan et al., 2001). Two older zircons (1620 and 1670 Ma) may have sources in the shield. In addition, the Poughquag Quartzite contains two zircons with dates of 643 and 547 Ma, corresponding to the ages of synrift igneous rocks.

In the central Appalachians, Eriksson et al. (2004) analyzed detrital zircons from the Hardyston Sandstone in eastern Pennsylvania, which is here interpreted to represent the rift-drift transition. The synrift Chestnut Hill Formation is sporadically distributed beneath the Hardyston Sandstone (Gates and Volkert, 2004). The Hardyston Sandstone is overlain by an extensive lower Cambrian-middle Ordovician carbonate succession that characterizes the passive margin. The detrital-zircon-age population is dominated by 950-1250 Ma (Grenville-age) zircons and a few grains with ages ranging from 550-650 Ma.

In central Alabama, the lower Cambrian Rome Formation is interpreted to be a late synrift sedimentary deposit (Thomas, 1991). Thomas et al. (2004b) analyzed detrital zircons from the Rome Formation for U-Pb ages. The Rome Formation is dominated by

Grenville-age (970-1240 Ma) and Granite-Rhyolite-age (1270-1540 Ma) detrital zircons, with subordinate zircon populations of 1610-1840 Ma (Yavapai-Mazatzal-Central Plains orogens), 1890-1970 Ma (Trans-Hudson/Penokean orogens), and 2310-2930 Ma (Superior craton).

Passive Margin

Within the passive-margin carbonate-shelf succession in Newfoundland, clastic interbeds (Hawke Bay Formation) contain detrital zircons with a wide range of ages (Cawood and Nemchin, 2001). In addition to Grenville-age (987-1109 Ma) zircons, the sandstone contains three older clusters of ages at 1229-1360, 1780-1860, and 2670-2790 Ma (Cawood and Nemchin, 2001). These are interpreted to represent older Grenville components, the mid-continent orogens, and the Superior province, respectively.

The early Cambrian Erwin Formation, of the Chilhowee Group in western North Carolina/Virginia, is interpreted to be a passive-margin sandstone (Eriksson et al., 2004). Detrital-zircon ages from the Erwin Formation have a 950-1250 Grenville-age mode, and several zircons with ages that range from 1300 to 1800 Ma (Eriksson et al., 2004).

In eastern Oklahoma and western Arkansas, an equivalent off-shelf deep-water passive-margin mud-dominated succession borders the Laurentian passive-margin carbonate-shelf facies. The off-shelf succession in the Ouachita thrust belt includes the Lower Ordovician Blakely Sandstone, a compositionally and texturally mature quartz-sand grain-flow deposit reworked from the shelf (Viele and Thomas, 1989). Detrital zircons from the Blakely Sandstone have dates that cluster at 1003-1188 and 2679-2722 Ma, as well as scattered dates of 1271, 1334, and 1744 Ma (Gleason et al., 2002). The detrital-zircon-age population is similar to that of the Rome Formation from central Alabama (Thomas et al., 2004b). The dates of 1271-1334 Ma are similar to dates of 1284-1407 Ma for granite boulders in the Blakely Sandstone (Bowring, 1984), suggesting that both boulder- and sand-sized detritus was supplied to the off-shelf slope from the Granite-Rhyolite province at the continental margin.

The basal passive-margin transgressive sandstones overlap the rifted margin and rest directly on Grenville-age basement rocks, consistent with the abundance of Grenville-age detrital zircons. The source of Grenville-age (950-1200 Ma) zircons within the overlying Paleozoic passive-margin sequence is problematic (Cawood and Nemchin, 2001; McLennan et al., 2001). The transgressive passive-margin strata progressively onlapped and covered the rim of Grenville rocks adjacent to the Laurentian margin (Rankin et al., 1989), and the entire Grenville province may have been covered by sediment and protected from erosion before deposition of the Grenville-age-zircon-bearing sandstones within both the shelf and off-shelf passive-margin successions. Because of the extensive stratigraphic cover, direct supply of detritus from exposed Grenville rocks to the shelf and shelf margin is unlikely, and some reworking of Grenville detritus is implied (Gleason et al., 2002). Transcontinental drainages distributed sediment from the Grenville orogen across the Laurentian continent in the late Proterozoic (Rainbird et al., 1997), suggesting a basis for recycling of Grenville detritus from intracratonic Proterozoic sedimentary basins to the passive-margin shelf adjacent to the Laurentian shield.

The older zircons in the passive-margin sandstones might represent direct sediment transport from primary sources in the mid-continent orogens and Superior province (e.g., Cawood and Nemchin, 2001). Alternatively, these zircons also might reflect recycling of older shield-derived detritus from Precambrian synrift sediment along the rifted margin.

Taconic Clastic Wedge

Synorogenic clastic-wedge deposits of Middle Ordovician to Silurian age document transport of detritus from the Taconic orogen into the foreland basin (e.g., King, 1951; Thomas, 1977). Detrital zircons in the Taconic synorogenic sediment from Newfoundland, southern New York, eastern Pennsylvania, and western Virginia are dominantly of Grenville age, and very rare zircons have ages that correspond to Iapetan synrift igneous rocks (Gray and Zeitler, 1997; McLennan et al., 2001; Cawood and Nemchin, 2001; Erikkson et al., 2004). Ordovician clastic-wedge sandstones in Newfoundland also have clusters of zircon ages at ~1800 Ma and 2700-2800 Ma (Cawood and Nemchin, 2001), whereas samples from western Virginia also include a few detrital zircons with ages of 1300-1800 Ma, 2008 Ma, 2525 Ma, and 3780 Ma (Eriksson et al., 2004). Although Taconic clastic-wedge deposits bracket the ages of volcanic, plutonic, and metamorphic rocks, they contain no detrital zircons that are from the contemporaneous hinterland igneous or metamorphic rocks, or from the Taconic-age volcanoes that are documented by numerous bentonite beds in the foreland stratigraphy.

In the Ouachita succession of off-shelf passive-margin strata, the laterally discontinuous Silurian Blaylock Sandstone is compositionally immature, suggesting a possible synorogenic provenance (Lowe, 1989; Viele and Thomas, 1989). Detrital zircons from the Blaylock Sandstone define clusters at 980-1186 and 1321-1409 Ma, and include a single grain at 486 ± 51 Ma (Gleason et al., 2002). Zircons with U-Pb dates of 453.1±1.3 and 454.5±0.5 Ma from K-bentonites preserved in passive margin strata (e.g., Tucker and McKerrow, 1995) also potentially could be recycled into late Paleozoic sedimentary deposits during Alleghanian exhumation. Therefore, incorporation of Ordovician-age zircon populations does not necessarily reflect incorporation of Taconian igneous rocks into the source region.

Acadian Clastic Wedge

Eastward thickening clastic deposits of Devonian to early Mississippian age in the Appalachian foreland reflect the exhumation and erosion of the Acadian orogen (e.g., King, 1951; Thomas, 1977). Detrital zircons from the Devonian Walton Formation of the Catskill Group (Acadian synorogenic clastic wedge) in southern New York have a very restricted distribution of ages, none of which are contemporaneous with the Acadian orogeny (McLennan et al., 2001). The detrital zircons define clusters of ages at 1018-1258 and 419-467 Ma. These ages document sedimentary detritus from thrust sheets of Grenville basement (or Grenville-derived sediment), as well as rocks that crystallized during the Taconic orogeny. Possible sources of Grenville-derived sediment include the synrift deposits and the Taconic clastic wedge. The lack of older (pre-Grenville) zircons suggests a localized source terrane either (1) with no contribution of sediment from synrift and passive-margin strata which commonly contain a wide distribution of older craton-detrital zircons (McLennan et al., 2001; Cawood and Nemchin, 2001), or (2) with local synrift deposits that were excluded from the shield-derived dispersal systems.

The Devonian Cloyd Conglomerate of western-central Virginia is also interpreted to be an Acadian synorogenic deposit. Similar to all other sedimentary deposits from the Appalachian foreland, there is a distinct mode in the detrital-zircon-age population from 950-1250 Ma, as well as a few grains with ages from 461 to 594 Ma, and a single grain with an age of 2178 Ma (Eriksson et al., 2004).

The detrital-zircon-age populations of late Proterozoic-middle Paleozoic strata along the eastern margin of Laurentia are dominated by a 950-1250 Ma mode from Grenville-age crust. The populations have a considerable range of ages, from 3780 Ma to 419 Ma. Widespread bentonite beds within the Appalachian foreland also contain zircons, the U-Pb ages of which confirm syndepositional volcanism during the Acadian (Devonian) orogenic events (Tucker et al., 1998). Recycling of these sedimentary deposits during the Alleghanian orogeny would obscure the identification of a single sedimentary source.

Summary of Potential Sources

Age identification of potential sources of sediment to the Appalachian basin reveals a significant overlap between potential sedimentary sources. Although the crystalline internides of the Appalachian orogen are comprised of rocks from Grenville and Paleozoic orogenic events, the detrital-zircon population from synrift, passive margin, and Paleozoic orogenic clastic wedges is much more diverse, containing detrital-zircon populations that also correspond to crust-forming events in Laurentia. This complication makes it very difficult to identify unique Laurentian sedimentary sources for the Alleghanian clastic wedge. Fortuitously, several peri-Gondwanan terranes were accreted to Laurentia prior to and during the late Paleozoic Alleghanian orogeny (e.g.

Rast et al., 1976; Horton et al., 1989; Hibbard, 2000), and they contain zircons with crystallization-age modes that contrast with those of Laurentia. If these terranes were a significant component of the Alleghanian tectonic load, it is expected that the detrital-zircon-age population will reflect provenance from these peri-Gondwanan terranes.

The subsequent chapters will focus on the application of U-Pb dating of detrital zircons to attempt to establish whether or not a temporal sequence of Alleghanian uplift and unroofing can be resolved in the sedimentary record in the Appalachian basin. These results will be combined with K/Ar ages of detrital white mica from the Appalachian basin to reconstruct the exhumational history of the Alleghanian orogenic belt. The ultimate intent is to improve our understanding of the tectonic history of the Alleghanian orogeny.

CHAPTER 2: DETRITAL ZIRCON EVIDENCE OF LAURENTIAN CRUSTAL DOMINANCE IN THE LOWER PENNSYLVANIAN DEPOSITS OF THE ALLEGHANIAN CLASTIC WEDGE IN EASTERN NORTH AMERICA

Introduction

The Alleghanian orogeny is an important component of the late Paleozoic construction of Pangea (Wilson, 1966; Bird and Dewey, 1970; Hatcher et al., 1989). Deformation, metamorphism, and igneous activity recorded along the Appalachian hinterland and the westward progradation of clastic wedges into the Appalachian foreland are collectively tied to a continental collision between the Gondwana and Laurentia in the late Paleozoic (reviewed in Hatcher et al., 1989). Among the possible components of the collision orogen are Laurentian and Gondwanan continental crust, synrift rocks of the Iapetan rifted margin of Laurentia, synorogenic rocks of the Taconic (Ordovician-Silurian) and Acadian (Devonian-Mississippian) orogenies, and Alleghanian synorogenic rocks. Erosion of the Alleghanian orogen supplied sediment for deposition in a foreland basin that formed in elastic response to tectonic loading of the Laurentian lithosphere (Fig. 1) (e.g. Jordan, 1981; 1995), and identity of the provenance of sandstones in the Appalachian foreland basin is a key link to the collisional anatomy of the Alleghanian orogen. To identify the various crustal components of the tectonic load, samples of early Pennsylvanian (early to middle Morrowan) sandstones and conglomerates from the Appalachian basin were collected along strike at locations most proximal to the Alleghanian crystalline thrust front for U-Pb age analysis of detrital zircons (Fig. 2). Because these early Pennsylvanian sandstones and conglomerates are preserved near the Alleghanian hinterland, they represent the direct possible sampling of the orogenic belt that was the source of sediment to the Appalachian basin. Paleocurrent data and

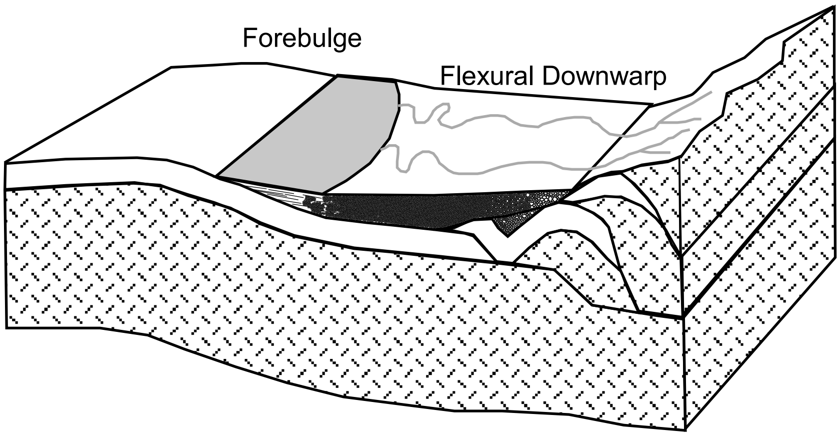

Figure 2.1. The composite orogen constitutes a gravitational load on the edge of the continental lithosphere. In elastic response to the tectonic load, the continental lithosphere produces a flexural downwarp, or foreland basin, providing accommodation space for sediment shed from the tectonic load. The sedimentary record in the foreland basin should reflect the progressive imbrication of crust on the margin.

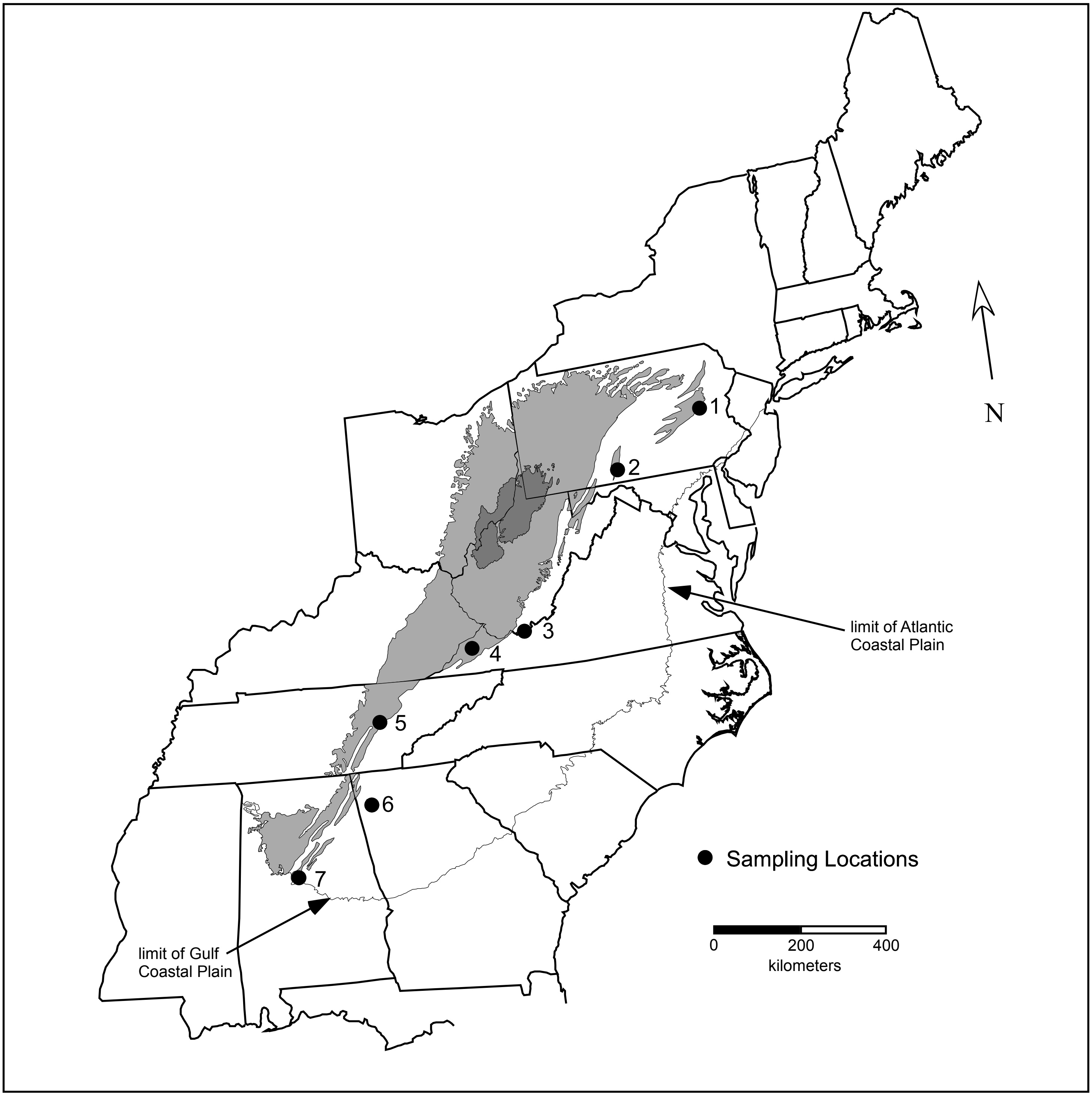

Figure 2.2. Map showing the locations of samples analyzed for detrital-zircon ages. The light gray shading represents the outcrop areas of sedimentary deposits of Pennsylvanian age in the Appalachian basin. The darker gray color corresponds to the outcrop area of Permian-age deposits. The samples are from: 1) Tumbling Run Member, Pottsville Formation; 2) Pottsville Formation; 3) Pocahontas Formation; 4) Lee Formation; 5) Sewanee Conglomerate, Crab Orchard Mountain Group; 6) Raccoon Mountain Formation; 7) Montevallo Coal Zone Member, Pottsville Formation.

progradation of facies indicate that the sediment was transported from the orogenic highlands (Thomas, 1977; Edmunds et al., 1979, 1999; Osborne, 1989, 1991).

Alleghanian hinterland

The presently exposed Appalachian hinterland is composed of crustal components which record two distinctly different petrogenetic histories: Laurentian and Gondwanan. The age of basement within the cratonic interior of Laurentia is comprised of the Archean Superior craton (2600-2800 Ma); the Paleoproterozoic Trans-Hudson and Penokean orogens (1800-1900 Ma); and the Mesoproterozoic Yavapai, Mazatzal, and Central Plains (1600-1800 Ma), Granite-Rhyolite (1300-1500 Ma), Keewanawan (1100 Ma), and Grenville provinces (950-1250 Ma) (Fig. 3) (Hoffman, 1989; Van Schmus et al., 1993). Gondwanan crust is genetically distinct from Laurentian crust because of the Late Proterozoic assembly of cratonic terranes to form the Gondwanan supercontinent during the Pan-African/Brasiliano orogeny (530-680 Ma) (Trompette, 2000). This event fused several cratons cored with Paleoproterozoic Trans-Amazonian/Eburnian (1950-2200 Ma) crust, and generated several continental-arc belts ranging in age from 680 to 530 Ma within Gondwana (Trompette, 2000). Fortuitously, 2.0-2.2 Ga crust is not well represented in Laurentia, with only a small continental fragment in the northwestern United States (Hoffman, 1989). In general, the late Proterozoic (580-680 Ma) and Paleoproterozoic (2000-2200 Ma) age crust is more characteristic of Gondwana (e.g. Mueller et al., 1994) than Laurentia. In contrast, Grenville-age (950-1250 Ma) continental crust underlies most of the Appalachian hinterland, thrust belt, and foreland basin (e.g., Hoffman, 1989).

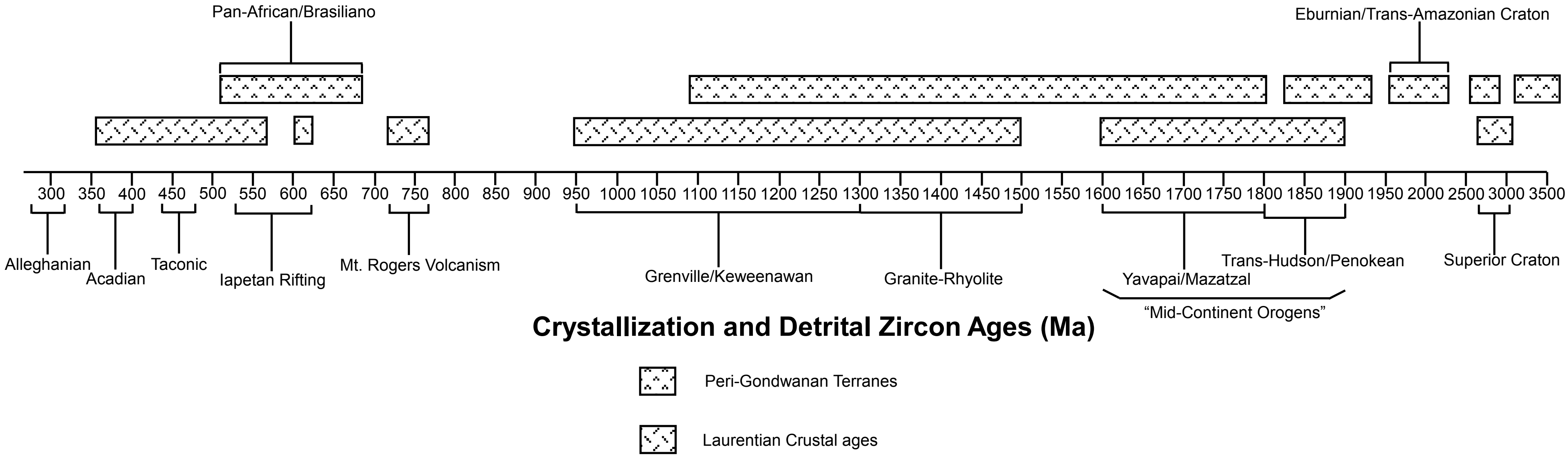

Figure 2.3. Bar graph representing composite zircon ages from peri-Gondwanan terranes (Samson et al., 2001; Coler and Samson, 2000; Wortman et al., 2000; Ingle-Jenkins et al., 1998; Mueller et al., 1994) and Laurentian crust (Hoffman, 1989; van Schmus et al., 1993; Aleinikoff et al., 1995).

The late Proterozoic-earliest Paleozoic Pan-African/Brasiliano orogenies that characterize Gondwana are coeval with diachronous rifting of eastern Laurentia and opening of the Iapetus Ocean. Iapetan rifting is documented by igneous rocks ranging in age from ~620 Ma in Newfoundland to 530 Ma along the Southern Oklahoma fault system (Hogan and Gilbert, 1998; Thomas et al., 2000; Cawood et al., 2001). An early phase of rifting along the southern Appalachian Laurentian margin has also been suggested to explain the limited distribution of 750-700 Ma Mount Rogers and related igneous rocks in western North Carolina and Virginia (Lukert and Banks, 1984; Goldberg et al., 1986; Su et al., 1994; Aleinikoff et al., 1995; Tollo and Hutson, 1996). Most of these igneous rocks have limited exposure and are unconformably overlain by lower Paleozoic strata, suggesting that they may not have supplied a substantial volume of sediment to the Appalachian basin. Possibly, detritus from these Neoproterozoic sources was incorporated in the lower Cambrian latest synrift and passive-margin strata, and subsequently recycled during the Alleghanian orogeny (e.g. Thomas et al., 2004a); however, the late Proterozoic Ocoee Supergroup, a late Proterozoic rift basin fill (Rast and Kohles, 1986) overlying the Neoproterozoic igneous rocks, does not contain Neoproterozoic (650-800 Ma) zircons (Bream, 2002). Some Neoproterozoic zircons have been found in synrift to passive-margin deposits in Newfoundland, southeastern New York, and eastern Pennsylvania (Cawood and Nemchin, 2001; McLennan et al., 2001; Eriksson et al., 2004).

Detrital-zircon age populations from synrift and passive-margin strata (Cawood and Nemchin, 2001; McLennan et al., 2001; Gleason et al., 2002; Eriksson et al, 2004) support the hypothesis that, during the late Proterozoic and early Paleozoic, drainage networks from the continental interior distributed sediment to the Laurentian margin of the Iapetus Ocean (outlined in Thomas et al., 2004a,b). The synrift and passive-margin deposits were likely exhumed during subsequent Paleozoic orogenic events along the Laurentian margin, serving as a source for recycled sediment that had an ultimate source in the interior of the Laurentian craton. Detrital zircons in Late Proterozoic and early-middle Paleozoic sediment have ultimate sources in the Archean Superior craton (2600-2800 Ma); the Paleoproterozoic Trans-Hudson and Penokean orogens (1800-1900 Ma); the Mesoproterozoic Yavapai, Mazatzal, and Central Plains (1600-1800 Ma), Granite-Rhyolite (1300-1500 Ma), Keewanawan (1100 Ma), and Grenville provinces (950-1250 Ma); and Iapetan synrift volcanics (620-530 Ma) (McLennan et al., 2001; Cawood and Nemchin, 2001; Gleason et al., 2002; Eriksson et al., 2004).

Prior to the late Paleozoic collision of Gondwana and Laurentia, several peri-Gondwanan terranes were accreted to Laurentia during the early and middle Paleozoic Taconic and Acadian orogenies (e.g. Rodgers, 1972; Rast et al., 1976; Williams and Hatcher, 1982; O'Hara and Gromet, 1985; Horton et al., 1989; Hibbard, 2000; Hibbard et al., 2003). The Taconic (ca. 490-440 Ma) and Acadian (ca. 390-350 Ma) orogenies included synorogenic emplacement of igneous rocks along the eastern margin of Laurentia. The superposition of Gondwanan terranes with respect to accreted Paleozoic arc terranes is cited as evidence of their exotic origin (Rast et al., 1976; Horton et al., 1989; Hibbard, 2000). Accreted Gondwanan terranes (Suwannee, Carolina, and Avalon) are also recognized because of provincial Gondwanan fauna (Secor et al., 1983) and crustal ages typical of Gondwana (Fig. 3) (Zartman and Hermes, 1987; Skehan and Rast, 1990; Mueller et al., 1994; Samson, 1995; Ingle-Jenkins et al., 1998; Heatherington and

Mueller, 1999; Coler and Samson, 2000; Wortman et al., 2000; Samson et al., 2001). The Goochland terrane, in eastern Virginia, hosts plutons of Neoproterozoic age (654 to 588 Ma) that are intruded into a Grenville-age basement (Owens and Tucker, 2003). Owens and Tucker (2003) interpret the ages to represent late Proterozoic rifting of a continental block of Laurentian origin; however, juxtaposition of the Goochland terrane east of the Ordovician-age Milton and Chopawamsic accreted arc terranes (Coler et al., 2000) and the timing of magmatism are consistent with a Gondwanan origin. Drilling in Chesapeake Bay east of the Chopawamsic arc terrane revealed a crystalline basement of peraluminous, weakly metamorphosed 614 +/- 9 Ma granite (Horton et al., 2001), likely indicating Gondwanan crust beneath the Atlantic Coastal Plain. The Neoproterozoic ages overlap with the Pan-African/Brasiliano orogenic event in Gondwana (Fig. 3).

The Alleghanian orogeny is commonly interpreted to be a result of orthogonal collision between Gondwana and Laurentia in the late Paleozoic (Wilson, 1966; Hatcher, 1972; 1987; Sinha and Zeitz, 1982; Rast, 1989). Orthogonal collision between Gondwana and Laurentia requires the destruction of oceanic lithosphere that would create an arc system and accretionary prism between Gondwana and Laurentia. Sinha and Zeitz (1982) proposed that several late Paleozoic granites along the eastern margin of Laurentia represent an Alleghanian (Hercynian) continental-margin arc system. An alternative model of formation of Pangea appeals to highly oblique collision between Laurentia and Gondwana along continental-scale dextral shear zones in the late Paleozoic (Gates et al., 1988; Shelley and Bossiere, 2000; Hatcher, 2002; Vai, 2003). The timing of displacement along several of these shear zones ranges from early Pennsylvanian to late

Permian (e.g., Stockey and Sutter, 1991; Maher et al., 1994; Pray et al., 1997; Wortman et al., 1998).

Detrital-zircon populations in the sedimentary fill of the foreland basin may help to distinguish between the alternatives for Alleghanian deformation. By late Paleozoic time, the eastern margin of Laurentia was rimmed by several peri-Gondwanan terranes (Fig. 4). A break-forward deformation sequence and a critical-taper model for the Appalachian hinterland (e.g., Hatcher and Williams, 1986) imply imbrication of Gondwanan crust onto the Laurentian continental margin, suggesting that detrital-zircon populations of the stratigraphically lowest synorogenic clastic deposits would contain ages that are characteristic of Gondwanan provenance. A highly oblique collision, as suggested by the dextral shear zones, implies uplifted crustal blocks of variable composition, possibly not including Gondwanan crust, thereby yielding a detrital-zircon population of characteristic Laurentian ages. A characterization of the provenance of the sediment in the Appalachian foreland basin, using detrital zircons to reveal the components of the tectonic load, will discriminate synorogenic igneous rocks, accreted Gondwanan terranes, and imbricated Laurentian continental crust.

Appalachian foreland

Prior to any manifestation of the Alleghanian orogeny, a middle-late Mississippian carbonate shelf extended across most of southeastern Laurentia (Fig. 5). Carbonate production was impeded before the end of the Mississippian by the progradation of mud, silt, and fine sand into the Appalachian foreland represented by the upper Mississippian Floyd and Parkwood formations in the Black Warrior basin, the

Figure 2.4. Map of the distribution of Gondwanan and Laurentian crust within the exposed crystalline Alleghanian hinterland (after Horton et al., 1989; Zartman et al., 1988). Recent work (e.g. Hibbard et al., 2003; Hatcher et al., 2002) suggests that more of the crust may be Gondwanan than previously thought.

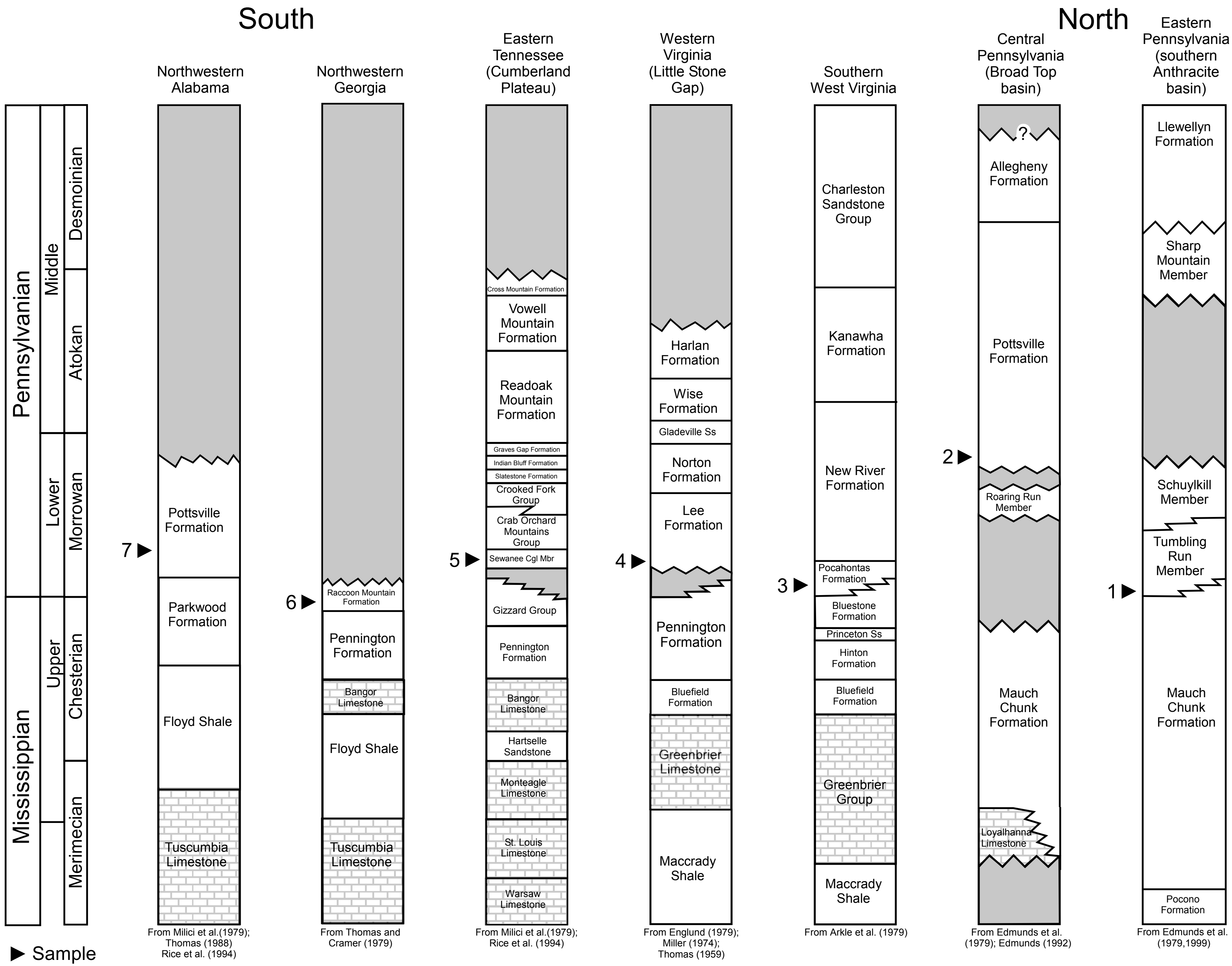

Figure 2.5. Stratigraphic columns of upper Mississippian to middle Pennsylvanian rocks in the Appalachian basin. Facies generally trend from a carbonate shelf in the middle to upper Mississippian to one dominated by clastic sediment influx, generally interpreted to mark the initiation of Alleghanian continental collision. Gray shading indicates lack of preserved sediment. Black triangles mark the stratigraphic unit sampled for detrital zircons. Numbers 1-7 correspond to sample locations shown in Figure 2.

Floyd and Pennington formations in the southern Appalachian basin, and the Bluefield and Mauch Chunk Formations in the central and northern Appalachian basin (Fig. 5) (Ferm, 1974).

In the Appalachian basin, the Pennsylvanian-Mississippian contact is characterized by an abrupt upward transition from marine-dominated shales, siltstones, and fine-grained sandstones to quartzose sandstones and a coal-bearing succession. This transition is generally interpreted to represent the cratonward progradation of a clastic wedge associated with the early stages of Alleghanian continental collision. Within the Appalachian basin, Thomas (1977) recognized two late Paleozoic clastic wedges: the Mauch Chunk-Pottsville in the central Appalachians (Pennsylvania salient) and the Pennington-Lee in the southern Appalachians (Tennessee salient), both attributed to the late Paleozoic Alleghanian orogeny. The southern part of the Pennington-Lee clastic wedge (Pennington-Raccoon Mountain Formations) in northwestern Georgia progrades westward over the late Mississippian Bangor Limestone in northeastern Alabama and southern Tennessee (Fig. 6) (Thomas, 1974). Farther west in the Black Warrior foreland basin and Appalachian thrust belt in northwestern Alabama and northeastern Mississippi, a separate clastic wedge (Floyd-Parkwood-Pottsville formations) progrades northeastward over the southwestern edge of the Bangor Limestone carbonate shelf, indicating a sediment supply from the Ouachita orogen on the southwest (Thomas, 1988). The northeast-prograding Ouachita clastic wedge merges with the southwest-prograding Pennington-Lee clastic wedge around the southeastern edge of the Bangor Limestone, and the Pennsylvanian components (both called Pottsville Formation) of the clastic wedges converge above the Mississippian Bangor Limestone in northeastern Alabama.

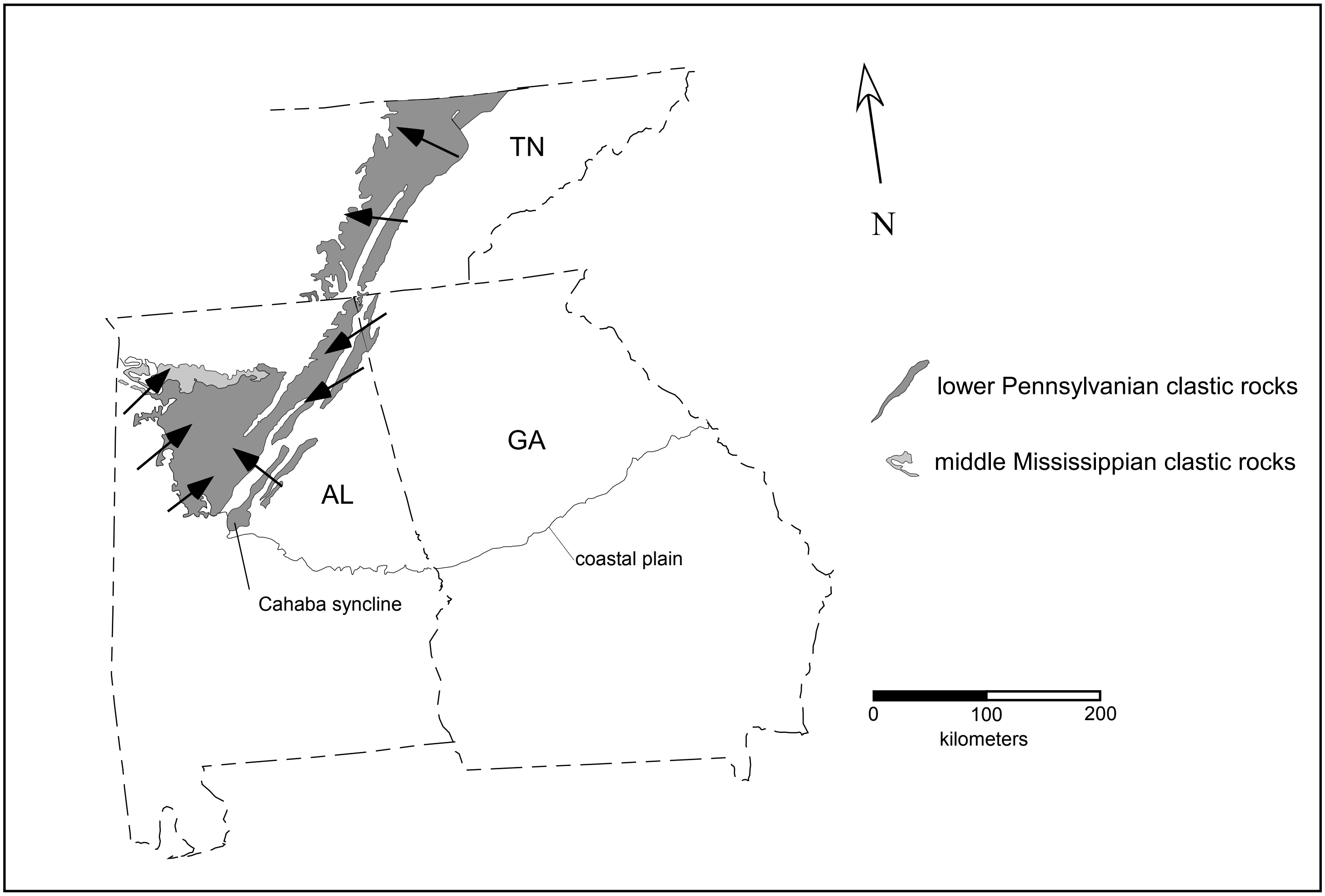

Figure 2.6. Sedimentary dispersal patterns for the late Paleozoic clastic wedges in the southern Appalachian and Black Warrior basins. In northeastern Alabama, southwestward progradation from the Pennington-Lee clastic wedge of Appalachian provenance intertongues with a northeastward prograding Ouachita clastic wedge. An early Pennsylvanian alluvial fan complex in central Alabama is recognized as a separate depositional system and is termed the Straven-Pottsville clastic wedge to differentiate it from the Ouachita and Pennington-Lee clastic wedges. Arrows denote the direction of sedimentary progradation of the clastic facies (Thomas,1977; Pashin, 1995; Mars and Thomas, 1999).

During deposition of the Pottsville Formation, an additional, less extensive clastic wedge (upper part of Pottsville Formation), prograded northwestward from a separate Alleghanian source on the southeast, and merged with the Ouachita and Pennington-Lee clastic wedges in the Appalachian thrust belt and southeastern Black Warrior basin in northern Alabama (Thomas and Schenk, 1988). The separate, northwest-prograding clastic wedge in the upper Pottsville of Alabama is herein called the Straven-Pottsville clastic wedge to distinguish it from the Pennington-Lee and Ouachita clastic wedges (Fig. 6).

Post-depositional deformation of the strata in the Appalachian basin implies that erosion and transport of late Mississippian through early Permian detritus from the Alleghanian orogen to the foreland basin was not significantly affected by topographic barriers resulting from leading thrust-belt structures. The timing of compressional events leading to the propagation of thrusts into the Alleghanian clastic wedges has been partially constrained by paleomagnetic studies of the Appalachian thrust belt. Diachronous basinwide remagnetization, linked to episodic fluid flow that precipitated new magnetic domains, affected the southern Appalachians earlier than the central Appalachians as indicated by Apparent Polar Wander path ages (Fig. 7) (Miller and Kent, 1988). Application of the Apparent Polar Wander path ages and paleomagnetic fold tests constrain the timing and sequence of folding in the central and southern Appalachian thrust belt. On the basis of the paleomagnetic data, the southern Appalachian thrust belt formed in the early-middle Permian (Miller and Kent, 1988), and the central Appalachian thrust belt formed in the middle-late Permian (Stamatakos et al., 1996). Additionally, the paleomagnetic data indicate that the Appalachian thrust belts formed in a break-forward

Figure 2.7. Map of Alleghanian foreland thrust belts in the Appalachians. The light gray lines show faults assigned to the southern Appalachian thrust belt, which propagated into the foreland earlier than did the central Appalachian thrust belt, represented by the black lines (Rodgers, 1967; Dean et al., 1988; Miller and Kent, 1988; Stamatakos et al., 1996). Ages were determined by matching magnetic inclination and declination of the secondarily remagnetizated beds to Apparent Polar Wander Path ages (Miller and Kent, 1988; Stamatakos et al., 1996).

sequence. Field observations, in particular the analysis of cross-cutting layer-parallel strain markers, such as stylolites, corroborate the diachroneity (Rodgers, 1967; Dean et al., 1988; Spraggins and Dunne, 2001). Concordance of stratigraphic thickness of the lower Permian Dunkard Group around low-amplitude folds in the distal parts of the thrust belt (Hatcher et al., 1989; Faill, 1998) and the observed lack of “growth strata” also support the paleomagnetic interpretation. In addition, late Pennsylvanian to late Permian recrystallization of illite throughout the Appalachian basin generally overlaps the paleomagnetic Apparent Polar Wander path age of the basin-wide fluid expulsion event (Elliott and Aronson, 1987, 1993; Aronson and Farrington, 2000). This suggests that the recrystallization of the illite and the remagnetization of the Paleozoic strata in the Appalachian basin are genetically linked to a basin-wide fluid-flow event that is interpreted to have been driven from enhanced topographic recharge (Bethke and Marshak, 1990) or from the migration of metamorphic brines (Oliver, 1986; Schedl, et al., 1992) as a result of collisional tectonism. Taken in sum, the existing data indicate that large-scale thrust faulting and folding of the proximal foreland basin post-dated deposition of the preserved synorogenic clastic sequence.

Samples: location, stratigraphy, depositional setting, petrography

Mauch Chunk-Pottsville Clastic Wedge

Tumbling Run Member, Pottsville Formation (eastern Pennsylvania) (1, Figs. 2 and 5)

The lower Morrowan Tumbling Run Member is the stratigraphically lowest of the three members of the lower-middle Pennsylvanian Pottsville Formation in eastern Pennsylvania (Fig. 5). On the basis of megafloral data, it is interpreted to be conformable with the underlying Mauch Chunk Formation (Read and Mamay, 1964; Meckel, 1967; Edmunds et al., 1999). It has a maximum thickness of 183 meters (600 feet) near sampling site 1 (Fig. 2) (Edmunds, 1988) and pinches out to the north and west (Meckel, 1967; Edmunds et al., 1999). The Tumbling Run Member fines upward to the siltstone- and coal-bearing Schuylkill Member (Fig. 5). These two members of the Pottsville Formation are disconformably overlain by the southwestward prograding Sharp Mountain Member (Edmunds, 1988; Robinson and Prave, 1995). Paleocurrents, percentage of lithic clasts, and particle size distributions all indicate a sediment supply for the Tumbling Run Member from the southeast (Meckel, 1967). The sample was collected on the east side of the northbound lanes of I-81 about 0.5 mile north of Delano, Pennsylvania, from the base of the Tumbling Run Member in the measured section of Inners and Lentz (1988). Volumetrically, the sample of the Tumbling Run Member is about 92% quartz and the remainder is lithic clasts of chloritic siltstone, phyllite, and rare hypabyssal intrusives, similar to the composition described by Meckel (1967). The quartz commonly exhibits undulose extinction and is partly polycrystalline. The clasts are matrix supported; grain-size distribution and composition of the clasts and matrix are summarized in Table 1.

Pottsville Formation (south-central Pennsylvania) (2, Figs. 2 and 5)

In the Broad Top basin in south-central Pennsylvania, the Pottsville Formation rests disconformably on the upper Mississippian Mauch Chunk Formation (Fig. 5). Above the unconformity, a thin marine shale (Roaring Run Member) of early-middle Morrowan age separates the coarser, quartz-rich sandstones of the Pottsville Formation

Sample Location	Petrologic constituents (%) quartz	feldspar	lithic frag.	Particle size (phi)	standard deviation (phi)	skewness	% clasts
Tumbling Run Member, Pottsville Formation, PA	92	0	8	-3.50	0.79	-0.05	60
Pottsville Formation, PA	98	2	0	-3.48	0.63	-0.09	70
Pocahontas Formation, WV	68	10	22	1.87	0.63	0.05	n/a
Lee Formation, VA	93	3	4	2.23	0.72	0.32	n/a
Raccoon Mountain Formation, GA	95	1	4	2.01	0.57	0.11	3
Montevallo coal zone, Pottsville Formation, AL	33	5	62	-3.30	1.08	-0.08	60

Table 2.1. Petrologic characteristics of basal Pennsylvanian sandstone samples from the Appalachian basin that were processed for detrital zircons. Petrographic classification of the sedimentary framework constituents was determined by point counts on >300 grains of thin sections stained for feldspars. Clasts were defined as any particle greater than 0.2 cm. Particle size, standard deviation, and skewness were determined using the techniques outlined in Blatt (1992).

from the red, hematitic siltstones and shales of the Mauch Chunk Formation (Edmunds, 1992). The basal Pennsylvanian sandstones and conglomerates within the Broad Top basin were deposited later than those of the Tumbling Run Member of eastern Pennsylvania (Edmunds et al., 1979; 1999). On the basis of stratigraphic correlation with cores from the western Pennsylvania coal fields, the Pottsville Formation is estimated to be about 61 meters (200 feet) thick in Broad Top basin. Paleogeographic and paleoenvironmental reconstructions for middle-late Morrowan indicate west-northwestward drainage in central and southwestern Pennsylvania (Edmunds et al., 1979; 1999).

The sample from this location was collected at the lowest conglomeratic bed 14.4 meters above the top of the Mauch Chunk Formation at the location of the measured section of Edmunds (1992). Palynological data from the underlying marine beds (Edmunds, 1992) and correlation with biostratigraphically dated strata in southwestern Pennsylvania (Edmunds et al., 1999) suggest a middle-late Morrowan depositional age for the sample.

Petrographically, the sample is a clast-supported quartz-pebble conglomerate, with rare lithic clasts of chloritic slate (Table 1). The sandy matrix is dominantly quartz (98%) with trace amounts of kaolinized feldspar (1-2%). The petrography of the Pottsville Formation in central Pennsylvania is subtly distinct from that of the Tumbling Run Member. Feldspar is lacking and lithic fragments are much more common in the Tumbling Run Member (Table 1). The quartz pebbles in the Pottsville of Broad Top basin exhibit features, such as incipient subgrain development, Fairburne lamellae, and undulose extinction reminiscent of tartan twinning in microcline, which are consistent with very high strain at low temperatures (e.g. Tullis et al., 1973). Randomly oriented micro-shear bands within the quartz pebbles do not cross or deform grain boundaries, indicating that the strain features are pre-depositional. Vitrinite reflectance values from this region (Levine and Davis, 1984) indicate that the post- depositional temperatures did not exceed 175 ^{o}C.

Pennington-Lee Clastic Wedge

Pocahontas Formation (southern West Virginia) (3, Figs. 2 and 5)

In southernmost West Virginia, the Mississippian-Pennsylvanian contact is within a conformable upward transition from gray mudstones and black clay shale of the upper Bluestone Formation to coal-bearing quartzose sandstones and shales of the Pocahontas Formation (Fig. 5) (Englund, 1974; Miller, 1974). The Pocahontas Formation is interpreted to represent a prograding fluvio-deltaic system with a sedimentary source from orogenic highlands to the east (Englund, 1974). It is a transitional facies between the underlying dominantly marine Bluestone Formation of the Mauch Chunk Group of Mississippian-age and the overlying fluvial-deltaic quartz arenites of the Pennsylvanian-age New River Formation (Arkle et al., 1979). The Pocahontas Formation has a maximum thickness of 230 meters (750 feet) in western Virginia (Englund, 1974). The sandstone within the Pocahontas Formation is in northeast-striking lobes that may represent barrier-island bars (Englund, 1974). The sandstones of the overlying New River Formation and laterally equivalent Lee Formation (in Virginia and Kentucky) are more quartzose, massive, and coarser grained than those of the Pocahontas Formation (Englund, 1979).

In southern West Virginia, a remnant of the basal sandstones of the Pocahontas Formation is preserved within the Hurricane Ridge syncline (Thomas, 1966). The sample location is on the east side of Hurricane Ridge, east of the Mercer County Airport, approximately 3.3 kilometers (two miles) northeast of Bluefield, Virginia, and 11.6 kilometers (seven miles) southwest of Princeton, West Virginia, on Airport Road. The sample was collected about 0.3 meters (one foot) above the Bluestone-Pocahontas contact as described in the measured section in Thomas (1959). The Pocahontas Formation is finer grained than the basal Pennsylvanian sandstones of northeastern Pennsylvania (Table 1). The sample analyzed for detrital-zircon age population is 68% quartz, 10% feldspar, and 22% lithic fragments (siltstone, slate, schist, gneiss, and chert) (Table 1).

Lee Formation (southwestern Virginia) (4, Figs. 2 and 5)

In western Virginia, the early Morrowan Lee Formation (Rice, 1994) marks the abrupt disconformable contact between marine-dominated shale and siltstone of the underlying Mississippian Pennington Formation and massive quartz-rich sandstone of the Lee Formation. To the northeast, the Late Mississippian/Early Pennsylvanian Pocahontas Formation fills the depositional gap between the Pennington and Lee Formations (Fig. 5) (Miller, 1974). The Lee Formation has a maximum thickness of about 520 meters (1700 feet) in western Virginia (Miller, 1974), along a linear northeast-southwest trend. The quartz arenites of the Lee Formation are generally interpreted as beach and barrier-island complexes associated with a near- shore marine environment (Englund and Delaney, 1966). The arenites typically grade upward from a massive quartz arenite into thin ripple-marked sandstone and shaly partings, silty lithic arenites, and thin coal (Miller, 1974). Low-angle planar cross-beds and ripple marks generally have a southwest dip

(Miller, 1974), suggesting dominant longshore transport parallel to the axis of the Appalachian basin.

The sample was collected along the northwestern limb of the Powell Valley anticline at the intersection of Virginia State Route 610 and 790 at Little Stone Gap, Virginia, in the Lee Formation in the measured section of Thomas (1959). The sample is medium grained, and is composed of 93% quartz, 3% feldspar, and 4% lithic fragments (siltstone, phyllite, schist, and chert) (Table 1).

Sewanee Conglomerate (eastern Tennessee) (5, Figs. 2 and 5)

The Sewanee Conglomerate is the basal Pennsylvanian stratigraphic unit in eastern Tennessee, and comprises the base of the Crab Orchard Mountains Group (Milici et al., 1979; Rice et al., 1994). The Sewanee Conglomerate has a maximum thickness of about 25 meters (82.5 feet) and thins to the northwest, implying a northwestward progradation from a southeastern source (Englund, 1974; Milici et al., 1979). It grades upward into the Whitwell Shale, an assemblage of shale and coal interpreted to represent a back-barrier system (Milici et al., 1979).

The sample was collected within the Cumberland Plateau on the eastern limb of the Sequatchee anticline in Bledsoe County, Tennessee, along Tennessee State Route 30 at the village of Emery Mill. Additional detail about the sample location is reported in Thomas et al. (2004a).

Raccoon Mountain Formation (northwestern Georgia) (6, Figs. 2 and 5)

The basal Pennsylvanian Raccoon Mountain Formation in northwestern Georgia lies conformably upon the shales and mudstones of the Mississippian Pennington

Formation (Fig. 5). The Raccoon Mountain Formation thickens to the southeast and is interpreted to represent a fluvial-deltaic succession prograding westward from an orogenic source on the east (Thomas and Cramer, 1979). The facies represented within the late Mississsippian-early Pennsylvanian sedimentary deposits of northwestern Georgia are interpreted to include a southwestward-prograding barrier-island complex that parallels the paleoshoreline (Thomas and Cramer, 1979).

The sample was collected from pebbly, bluff-forming sandstone along the southeastern crest of Rock Mountain, a small synclinal basin within the Floyd synclinorium of northwestern Georgia (Thomas and Cramer, 1979). Although there is no paleontological constraint on the depositional age of the sandstone at Rock Mountain, it is presumed to be of Pennsylvanian age because of the lithologic similarity of the underlying shales to those of the Pennington Formation.

The sample is composed of fairly well-sorted sand with a few pebbles as much as 1 cm in diameter (Table 1). The framework grains are 95% quartz, 4% lithic fragments (siltstone, slate, schist, and chert), and 1% feldspar.

Straven-Pottsville Clastic Wedge

Montevallo Coal Zone, Pottsville Formation (central Alabama) (7, Figs. 2 and 5)

The Pottsville Formation in central Alabama lies conformably above the upper Mississippian-lower Pennsylvanian Parkwood Formation (Fig. 5). Within the Cahaba synclinorium of central Alabama, it is composed of interbedded coarse lithic conglomerates, sandstones, shales, and coal. Palynological and megafloral data indicate that the Pottsville is temporally equivalent to the Pocahontas and New River Formations of West Virginia (Smith, 1979). Within the Cahaba synclinorium, the Pottsville Formation is estimated to be >1525 meters (>5000 feet) thick (Pashin et al., 1995).

The petrology and distribution of the Straven Conglomerate Member of the Pottsville Formation has been studied in detail (Butts, 1926; Osborne, 1988). The Straven Conglomerate lies approximately 250 meters (825 feet) below the conglomerates of the Montevallo coal zone (Osborne, 1988; 1991). Clast imbrication orientations and mean particle size distributions suggest a sedimentary source to the southeast (Osborne, 1988), consistent with a supply from the nascent Alleghanian orogen. Osborne (1988, 1991) concluded that most of the sedimentary lithic clasts within the Straven Conglomerate are similar to rocks within the nearby lower-middle Paleozoic succession in the Appalachian thrust belt and Talledega Slate belt.

The Montevallo coal zone conglomerates share most of the same petrologic characteristics as the Straven Conglomerate. Both are dominated by recycled sedimentary and metasedimentary rock clasts (Table 1). Volcanic clasts have been observed within the Straven Conglomerate (Osborne, 1988, 1991), and a few have been observed in outcrop of the Montevallo coal zone (D.M. Surles, pers. comm.). The textural and compositional immaturity of the Straven and Montevallo conglomerates requires a very proximal source.

The sample was collected within the Cahaba synclinorium at Falling Rock Falls, Alabama. The location is in a measured section in Pashin et al. (1995), and is within the Montevallo coal zone. The matrix-supported conglomerate is comprised chiefly of lithic clasts > 4 cm in diameter in the following proportions: 30% quartzite, 18% sandstone, 18% chert, 30% quartz, and 4% igneous (rhyolite). The sandy matrix is 33% quartz, 4%

feldspar, and 62% lithic fragments (sandstone, siltstone, slate, phyllite, and chert). A sample of conglomerate, including the clasts and matrix, was processed together for detrital zircons.

Methods

Prior to crushing, the samples were washed to remove any potential contamination from eolian dust. After each sample was crushed, all components of the jaw crusher and disk mill were cleaned thoroughly to avoid any potential cross-contamination between samples. The resulting fine sand was separated into populations by density using a Wilfley table. The two densest fractions were collected and dried, and ferromagnetic phases were removed using a hand magnet. The remaining nonmagnetic dense mineral fraction was poured into a separatory funnel with doubly filtered pure 1, 1, 2, 2 tetrabromoethane (density =2.967 grams/milliliter). The dense split was collected, dried, and sieved into a size fraction (106-150 microns) that is recommended for determination of provenance sensistive heavy mineral ratios (Morton and Hallsworth, 1994). The 106-150 micron size fraction was hand-picked for detrital zircons, which were subsequently divided into populations based on color (yellow, pink, opaque, etc.) and morphology (faceted or round), similar to the procedure outlined in Thomas et al. (2004a). Representative samples from each of the color/morphologic categories were included in each composite sample. The purpose of this approach was to try to identify and include all of the potential sources. The detrital zircon populations were embedded in an epoxy mount, which was polished to expose the zircon cores.

U-Pb geochronology was conducted at the University of Arizona Geochronological Laboratory using a Micromass Isoprobe ICPMS equipped with nine

Faraday collectors, an axial Daly detector, and four ion-counting channels. The Isoprobe is connected to a New Wave DUV 193 laser ablation system, which has an emission wavelength of 193 nm. The analyses were conducted on 50-micron spots with an output energy of ~32 mJ and a repetition rate of 8 hz. Each analysis consisted of one 20-second integration on backgrounds (on peak centers with no laser firing) and twenty 1-second integrations on peaks with the laser firing. The depth of each ablation pit is ~20 µm. The collector configuration allows simultaneous measurement of ^{204}Pb in a secondary electron multiplier while ^{206}Pb, ^{207}Pb, ^{208}Pb, ^{232}Th, and ^{238}U are measured with Faraday detectors. All analyses were conducted in static mode.

Inter-element fractionation during the analysis was monitored by analyzing fragments of a large concordant zircon crystal that has a known (ID-TIMS) age of 564 ± 4 Ma (G.E. Gehrels, unpublished data). This reference zircon was analyzed once for every four unknowns. The calibration correction contributes ~3% (2-sigma) systematic uncertainty to both ^{206}Pb / ^{238}U and $^{206}Pb/^{207}Pb$ ages. The isotopic ratios are also corrected for common Pb using the measured ^{204}Pb, assuming an initial Pb composition according to Stacey and Kramers (1975) and uncertainties of 1.0 and 0.3, respectively, for $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$. The reported ages are based primarily on $^{206}Pb/^{207}Pb$ ratios for >1200 Ma grains and $^{206}Pb/^{238}U$ ratios for <1200 Ma grains (Appendix A). This approach is used because of the small quantity of radiogenic ^{207}Pb in young zircons. At about 1200 Ma, enough ^{207}Pb has been produced by decay of ^{235}U to allow for high precision measurements of $^{207}Pb/^{206}Pb$. Discordance in young grains is addressed by looking for clustering of ages in the detrital-zircon populations. In this paper, single-age populations are reported, but their significance is not emphasized because there is no independent method of determining discordance (hence the geologic significance) of a single $^{238}U/^{206}Pb$ age. One sample, the Sewanee Conglomerate from eastern Tennessee, was analyzed and reported previously (Thomas et al., 2004a).

Because the cores of zircons were analyzed, this method may misrepresent a potential source where incorporation of xenocrystic zircons into zirconium-oversaturated melts has occurred (Watson and Harrison, 1983). Detailed studies of several southern Appalachian granites reveal that some of the melts were Zr oversaturated, preventing inherited zircons from the surrounding country rock to be resorbed into the melt (Miller et al., 2000). Although zircons are used as a proxy for sedimentary source, the question remains as to whether the dominant sediment-making mineral (e.g. quartz) was derived from a different source than the zircon. Unfortunately linking the crystallization ages of major rock-forming constituent minerals to the U-Pb ages of the detrital zircons would be extremely difficult.

Results

The detrital zircons from the Alleghanian clastic wedge represent a wide range of crystallization ages, and seemingly a wide variety of sources (Fig. 8, Appendix A). Recycling of older sedimentary units of the late Proterozoic-early Paleozoic rift and passive margin accounts for the presence of the older zircon populations and is consistent with an orogenic source along the continental margin (Thomas et al., 2004a). Mesoproterozoic-age detrital zircons could come from the recycled sedimentary deposits or exhumation of Grenville-age basement. Because of chemical and mechanical resilience (Morton and Hallsworth, 1994), zircon is concentrated in the insoluble residuum that constitutes sedimentary rocks. Recycling of sedimentary rocks, therefore,

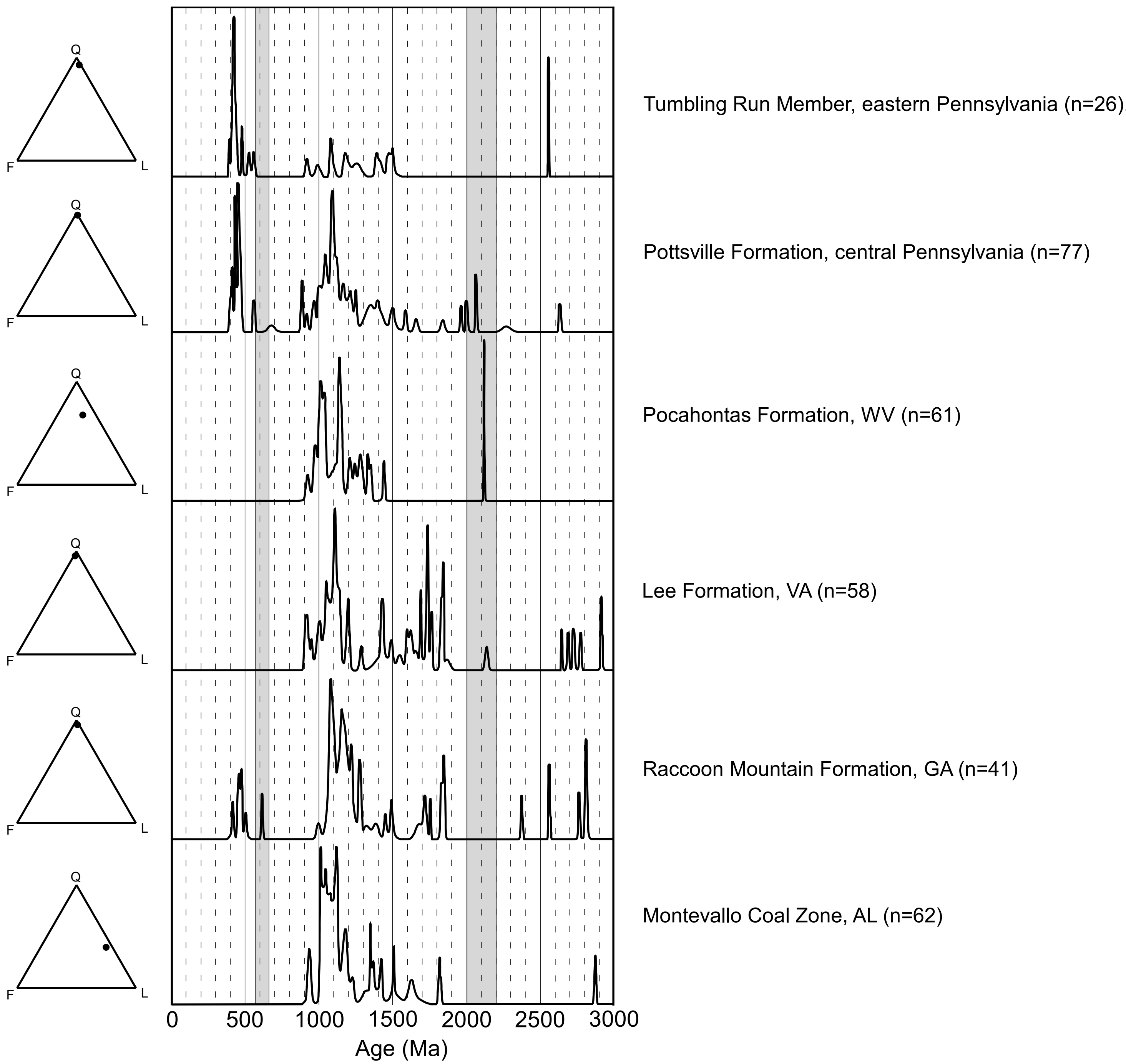

Figure 2.8. Age-probability plots of new detrital zircon ages from six basal Pennsylvanian sandstones in the Appalachian basin (see Appendix A). QFL plots for each sample are displayed to the left of each age-probability plot. The petrologic constituents of each sample are listed in Table 1. Gray bars represent crustal ages that are typical of Gondwanan crust. Detrital-zircon-age populations appear to have a more restricted range of ages in samples of litharenite composition.

results in a disproportionately large contribution of zircons to the detritus that is transported to the sedimentary basin. Recycling of sedimentary deposits may account for the large proportion of Mesoproterozoic, Paleoproterozoic, and Archean detrital zircons. Paleozoic detrital-zircon ages could be derived from igneous rocks associated with the Paleozoic Taconic (490-440 Ma) and Acadian (390-350 Ma) orogenies.

Most detrital zircons in the Alleghanian clastic wedge are of Laurentian affinity (Fig. 8); however, a few detrital zircons have crystallization ages (620-670 Ma; 2000-2300 Ma) that do not correspond to any substantial crustal component within Laurentia (Fig. 3) (Mueller et al., 1994). The rare ~2000-2300 Ma zircons correspond in age to the Trans-Amazonian/Eburnian events of Gondwana. Documentation of Trans-Amazonian-age detrital zircons within the cover sequences of peri-Gondwanan Carolina, Suwannee, and Avalon accreted terranes (e.g. Mueller et al., 1994) along the eastern margin of North America indicates a logical source for these age populations. The 620-670 Ma zircons overlap with the ages of volcanic and plutonic units within the peri-Gondwanan accreted terranes (Zartman and Hermes, 1987; Mueller et al., 1994; Samson, 1995; Ingle-Jenkins et al., 1998; Coler and Samson, 2000; Wortman et al., 2000; Samson et al., 2001; Eriksson et al., 2004) and add additional support to the hypothesis that they were integrated into the drainage network.

The detrital-zircon age populations show an apparent relationship to the composition of the Pennsylvanian clastic deposits (Fig. 8). The detrital-zircon age populations appear to be more restricted in samples with a lower proportion of quartz (e.g., Pocahontas Formation of West Virginia and Montevallo Coal Zone of Alabama). This relationship may reflect the influence of recycling of pre-orogenic sedimentary deposits from the Laurentian margin. Second-cycle deposits likely have a smaller proportion of lithic fragments and feldspar compared to first-cycle deposits, because of chemical dissolution of feldspar and mechanical breakdown of lithic fragments (Johnsson, 1993). Recycling of sedimentary deposits would result in an increase in the proportion of quartz and in the diversity of detrital-zircon ages.

Tumbling Run Member, Pottsville Formation (eastern Pennsylvania)

The Tumbling Run Member of the Pottsville Formation in eastern Pennsylvania is interpreted to have a source in the metamorphic rocks of the Piedmont to the southeast. Early-middle Paleozoic detrital zircon ages (389-476 Ma) from the Tumbling Run Member (Fig. 8) coincide with the age of arc magmatism, crustal thickening, and amphibolite-granulite metamorphism in the Wissihickon Schist and Wilmington Complex of southeastern Pennsylvania and northern Delaware (Bosbyshell et al., 1998; 2001). The other detrital zircon ages (522-555, 915-1498, 2564 Ma) have counterparts within Laurentia (Fig. 8). None of the detrital-zircon ages suggest derivation from a peri-Gondwanan source.

Pottsville Formation (south-central Pennsylvania)

Most of the detrital zircons have ages that correspond to crustal ages of Laurentia (389-548, 874-1830, 2628 Ma); however, some detrital zircon ages (666, 1989, 2054, 2262 Ma) suggest derivation from a peri-Gondwanan source (Fig.8). The supply of sediment for the lower Pennsylvanian sandstones in south-central Pennsylvania was from the southeast, presumably from the Appalachian Piedmont.

Pocahontas Formation (southern West Virginia)

The Pocahontas Formation does not contain zircons younger than ~900 Ma, and most of the zircon ages range from 900 to 1500 Ma, corresponding to parts of Laurentian crust. A single zircon with an age of ~2112 +/- 2 Ma corresponds in age to Trans-Amazonian/Eburnian-age crust, suggesting a possible contribution from Gondwanan terranes.

The substantial proportion of Grenville-age zircons and lack of younger Paleozoic zircon populations, along with a relatively high proportion of detrital feldspar, support a primary source from Grenville basement. A large percentage of the lithic fragments is chloritic slate and phyllite, but these rocks may be too fine grained to contribute zircons of the size (106-150 µm) analyzed.

Lee Formation (western Virginia)

Although deposition of the Lee Formation of western Virginia was nearly coeval with the Pocahontas Formation of southern West Virginia, the ages of detrital zircons from these samples contrast significantly. The Lee Formation contains a significant number (39%) of detrital zircons that range in age from 1400 to 2925 Ma with clusters at 1400-1490 Ma, 1590-1690 Ma, 1725-1760 Ma, 1820-1860 Ma, and 2650-2925 Ma. These older detrital zircons correspond to Laurentian cratonic sources, consistent with a general model for recycling of synrift and passive-margin sedimentary sources for these sandstones (e.g. Thomas et al., 2004a). Only a single 2134 +/- 10 Ma zircon in the sample suggests the possibility of a peri-Gondwanan source.

Petrographically, the Lee Formation has a much higher proportion of quartz than does the Pocahontas Formation, suggesting greater recycling of pre-orogenic strata into the Lee Formation. In contrast, a much higher proportion of feldspar suggests incorporation of a primary source in the Pocahontas Formation in the Hurricane Ridge syncline of southern West Virginia.

Sewanee Conglomerate (eastern Tennessee)

The detrital zircon population from this sample was reported previously in Thomas et al. (2004a), but is reviewed here for comparison with sandstones of similar depositional age along the Appalachian basin. The detrital-zircon dates from the Sewanee Conglomerate cluster at 385-465 Ma, 900-1350 Ma, 1700-1820 Ma, and 2690-2755 Ma; in addition, dates are scattered sparsely between 1460 and 1540 Ma. All of these ages correspond to crust-forming events in Laurentia. A single detrital-zircon age of 2122 +/- 10 Ma reflects the possible incorporation of a peri-Gondwanan crustal component.

The greater abundance of Paleozoic-age zircons in the Sewanee Conglomerate than in the Pocahontas and Lee Formations of West Virginia and Virginia, respectively, reflects exhumation and incorporation of Taconian and Acadian orogenic complexes that are exposed within the Piedmont and Blue Ridge physiographic provinces.

Raccoon Mountain Formation (northeastern Georgia)

The Raccoon Mountain Formation of northwestern Georgia contains detrital zircons of early Paleozoic age (450-499 Ma), Mesoproterozoic to Paleoproterozoic age (991-1848 Ma), and Archean age (2558-2815 Ma), all consistent with Laurentian sources.

Two zircons have ages that are typical of peri-Gondwanan crust (612 Ma, 2369 Ma). The detrital-zircon-age population is similar to that of the Sewanee Conglomerate.

Montevallo Coal Zone, Pottsville Formation (central Alabama)

The population of detrital-zircon ages from the Montevallo Coal Zone sample is wholly Laurentian in character; the ages span the Meso- and Paleoproterozoic (931-1817 Ma) and include a single Archean grain (2880 +/- 6 Ma). No zircons younger than Grenville were found. The clasts in the Straven Conglomerate show a close correspondence to rocks in the proximal Talledega slate belt and the southeastern thrust belt (Osborne, 1988; 1991). By inference, the cover sequence of the Talladega Slate belt is of Laurentian affinity, and may be part of an internally drained pull-apart basin (Ferrill and Thomas, 1988), a thick-skinned thrust block, or backarc extensional basin (Tull, 1998). The detrital-zircon population is similar to those of finer grained deposits (Pocahontas, Lee, and Raccoon Mountain formations) in the Appalachian basin. The Pottsville Formation in central Alabama may be an analog for unpreserved more proximal deposits all along the Appalachian thrust front.

Discussion

Destruction of oceanic lithosphere between eastern Laurentia and western Gondwana would require the existence of an Alleghanian arc system. Several late Paleozoic granites along the eastern margin of Laurentia have been interpreted to represent an Alleghanian arc system that developed prior to continental collision (Sinha and Zeitz, 1982). The ages were based largely on whole rock Rb-Sr analyses (e.g., Fullagar, 1971; Fullagar and Butler, 1979). A more recent re-evaluation of the age of these plutons by high precision U-Pb dating of zircon reveals that the emplacement of granites was restricted in time at about 325-295 Ma (Heatherington, 1998; Heatherington and Mueller, 1999; Samson and Secor, 2000; Schneider and Samson, 2001; Samson, 2001; Miller et al., in review), not 330-265 Ma as previously inferred (Sinha et al., 1989). In addition, geochemical constraints on the petrogenetic source of these granites reveal no mantle-derived component, indicating that they are not part of an Alleghanian arc complex (Samson et al., 1995; Coler et al., 1997). The emplacement of late Paleozoic granites may be tied to crustal-scale transpressional exhumation and advection of hot crust along dextral shear zones in the Appalachian hinterland (e.g., Gates et al., 1988; Brown and Solar, 1999; Teyssier and Whitney, 2002; Weinberg et al., 2004). The timing of dextral displacement along many of these shear zones, such as the Modoc (Sacks et al., 1993), Brevard (Stockey and Sutter, 1991), Central Piedmont (Dennis and Wright, 1995), Hyco (Wortman et al., 1998), Brookneal (Gates et al., 1986), Pleasant Grove (Krol et al., 1999), and Huntingdon/Cream Valley (G.Solar, pers. comm.) overlaps with the deposition of the clastic wedge within the Appalachian basin (~322-290 Ma). This dextral shearing event is also recognized to predate the brittle, west-vergent, foreland thrusting associated with the continental collision event (Maher, 1987; Hatcher, 2001).

The predominance of detrital zircons of Laurentian ages within basal Pennsylvanian sandstones suggests that the early phases of Alleghanian deformation did not involve Gondwanan crust along the continental periphery. The early Pennsylvanian displacement of dextral shear zones within the Appalachian hinterland suggests that this phase of deformation was responsible for the development of early Pennsylvanian topography. The composition of exhumed crust was of Laurentian affinity, some of which had detrital-zircon ages that corresponded to the cratonic interior of North America. The detrital zircon ages from basal Pennsylvanian sandstones reveal that Alleghanian deformation did not result in the incorporation of substantial exotic, non-Laurentian crust into the Appalachian sedimentary basin.

The collisional history of the Pangean assembly is also preserved in Pennsylvanian-age deposits on Gondwana (van Houten, 1976). Scattered late Paleozoic strata in Morocco and Algeria, which constituted part of the Gondwanan conjugate margin during the Alleghanian orogeny, are the stratigraphic complement to the Pennsylvanian-age deposits in the Appalachian basin. Polymictic conglomerates ranging in thickness from tens to hundreds of meters range in depositional age from late Atokan (Westphalian C) to Wolfcampian (Autunian) (van Houten, 1976). It is notable that these deposits are younger than the basal Morrowan clastic wedge in the Appalachian basin, further supporting the interpretation that continental collision post-dated transpressional uplift.

Invoking a transpressional orogen as the source for basal Pennsylvanian sandstones may help to explain some of the peculiar stratigraphic characteristics within the Appalachian basin, such as the lack of an initial marine deepwater deposit associated with loading of the Laurentian margin, as that in the Ouachita foredeep (Viele and Thomas, 1989). If the composition of the early Pennsylvanian Alleghanian orogeny is reflected in the detrital-zircon populations preserved in the Appalachian basin, then the kinematic history of an oblique collision must have resulted in the formation of orogenic highlands comprised of Laurentian crust.

Conclusions

Detrital-zircon age populations of basal Pennsylvanian sandstones from the Appalachian basin reveal that the supply of sediment for Alleghanian clastic-wedge sandstones is dominantly from Laurentian sources. This result is independent of grain size, depositional environment, or location along strike of the Appalachian basin. The documentation of dextral shearing along the eastern Laurentian margin coeval with the deposition of clastic detritus in the Appalachian basin suggests that transpressionally uplifted crustal blocks supplied sediment to the foreland.

CHAPTER 3: THE TECTONIC EVOLUTION OF THE ALLEGHANIAN OROGEN AS INTERPRETED FROM RADIOMETRIC SEDIMENTARY PROVENANCE PROXIES

Introduction

The late Paleozoic Alleghanian orogeny is interpreted to represent a collision between the continents of Laurentia and Gondwana to form the supercontinent Pangea (Fig. 1). Early interpretations of the Alleghanian orogeny appealed to an orthogonal plate collision between eastern Laurentia and western Gondwana during the late Pennsylvanian. This intuitive model explained the westward vergence of Appalachian thrust sheets away from the crystalline collisional front (e.g. Wilson, 1966; Hatcher, 1972, 1978, 1987; Lefort and Van der Voo, 1981; Rast, 1989). An orthogonal plate collision between Gondwana and Laurentia would have resulted in the destruction of oceanic lithosphere along one of the plate margins. In accordance with this hypothesis, Sinha and Zeitz (1982) proposed that several late Paleozoic granites in the southern Appalachian Piedmont (Fig. 2) represent a continental arc system associated with subduction beneath the present Atlantic margin. Deep seismic-reflection profiles acquired by the COCORP program also reveal subhorizontal reflectors that were interpreted to be crustal-scale thrust faults projecting to the surface at the location of the Brevard and Augusta fault zones (Cook et al., 1981). These faults were interpreted to have translated the Alleghanian hinterland westward toward the Laurentian continental interior (Cook et al., 1981).

Several lines of evidence recently have cast doubt on the validity of an orthogonal plate collision. Detailed field work in the crystalline internides of the Appalachians reveals that many of the major faults interpreted to be associated with thrusting the

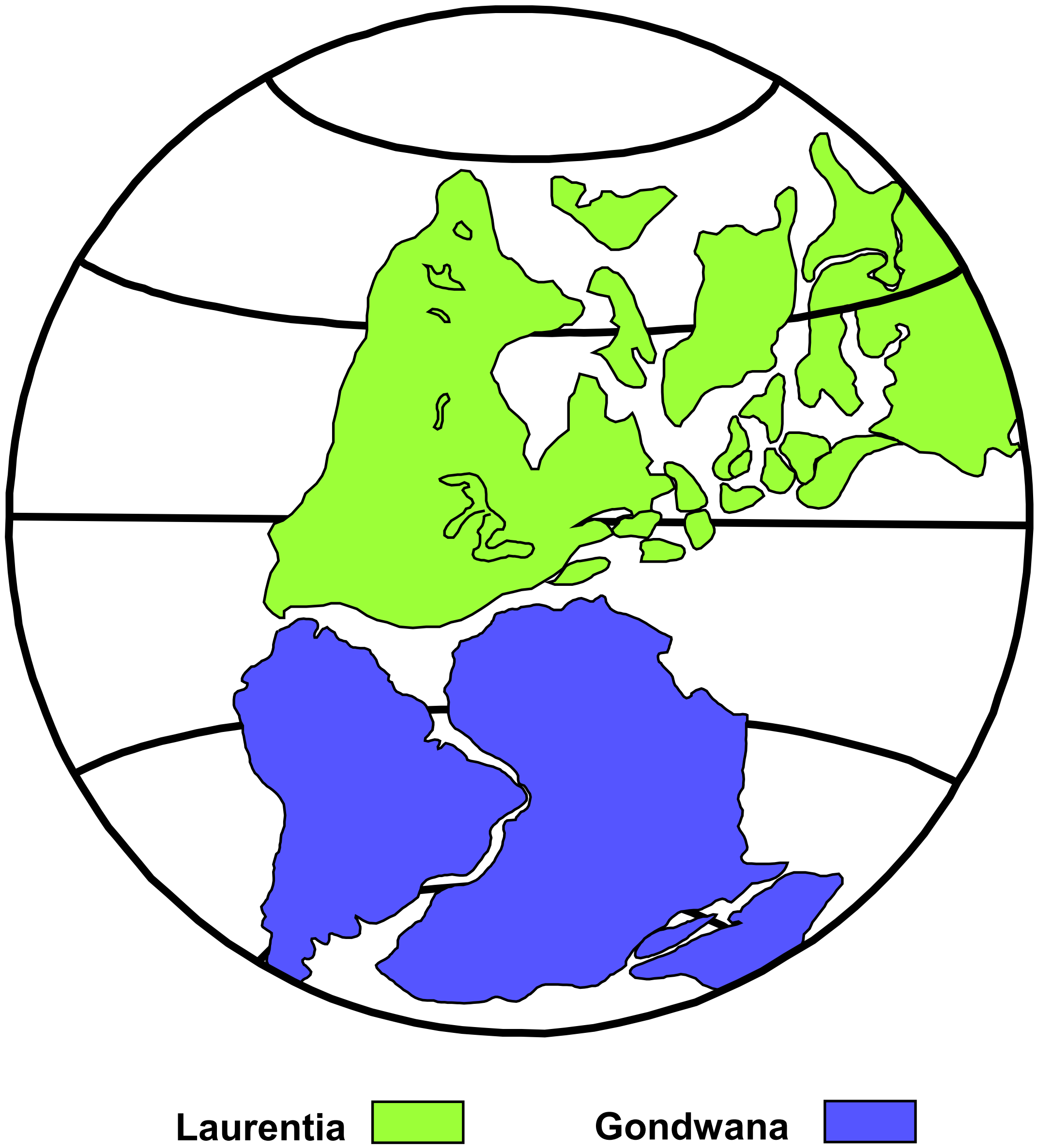

Figure 3.1. Global distribution of continents at approximately 300 Ma forming the supercontinent Pangea. Modified from Van der Voo and Torsvik (2001).

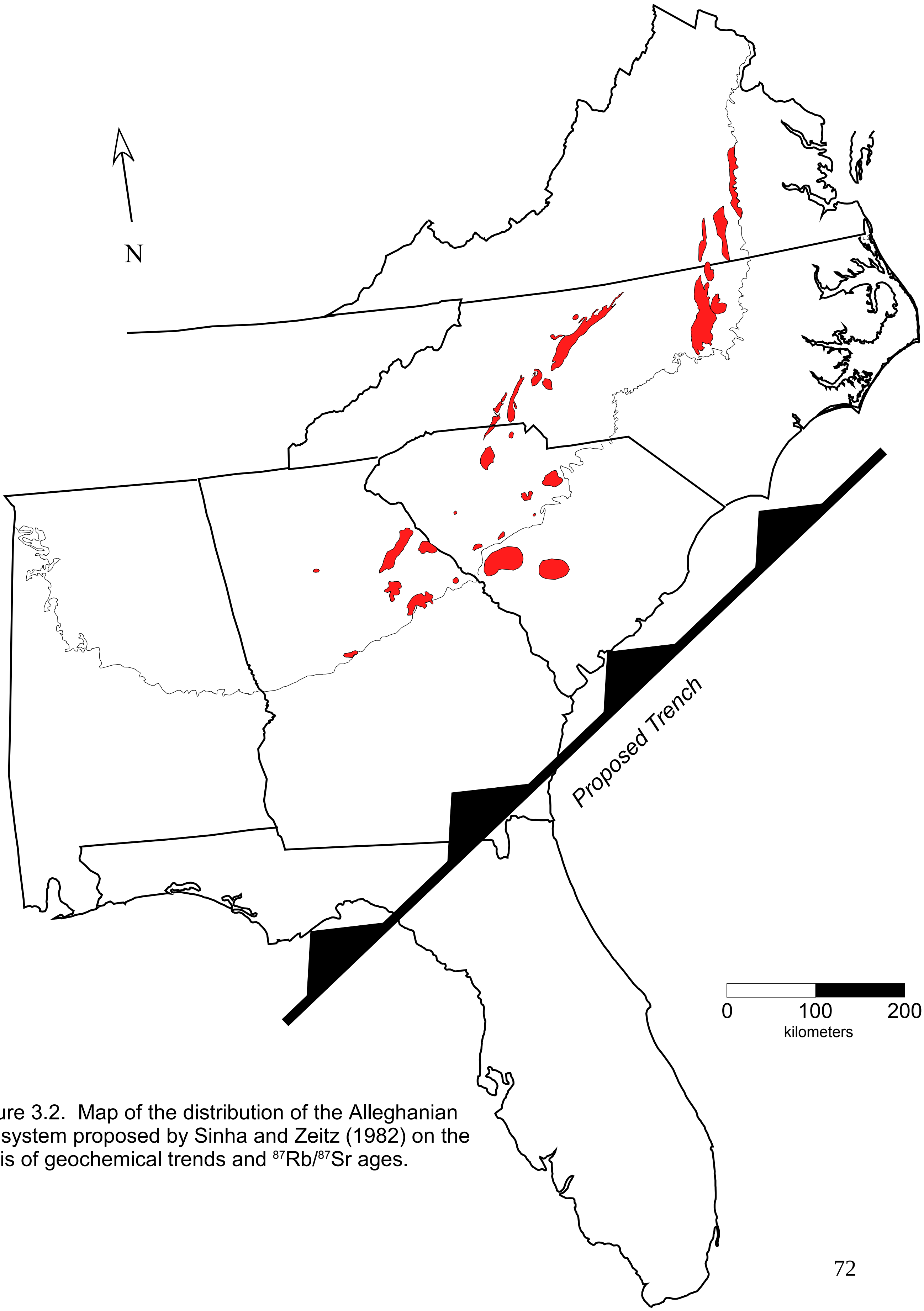

Figure 3.2. Map of the distribution of the Alleghanian arc system proposed by Sinha and Zeitz (1982) on the basis of geochemical trends and $^{87}Rb/^{87}Sr$ ages.

Alleghanian allochthon onto Laurentia, such as the Brevard and Augusta faults (Cook et al., 1981), actually exhibit dominantly dextral and/or normal displacement (Fig. 3) (Edelman et al., 1987; Maher et al., 1994). Geochemical studies of the late Paleozoic granites show that they do not have the attributes of a continental arc system (Samson et al., 1995; Coler et al., 1997). Both discoveries severely complicate models of orthogonal convergence and collision between Gondwana and Laurentia. As a result, many now infer that the plate collision responsible for the formation of the Pangean supercontinent was highly oblique and that the Alleghanian deformation was primarily translational in nature (e.g., Gates et al., 1986, 1988; Rodgers, 1987; Hatcher, 2002; Vai, 2003; Engelder, 2004).

The generally westward translation of thrust sheets into the Appalachian foreland, however, requires that a west-verging compressional component accompanied oblique plate collision. It has also been recognized that deformation of the Appalachian foreland did not occur during a single phase of plate collision. Several lines of evidence indicate that the thrust belt in the southern Appalachians formed prior to that in the central Appalachians (Fig. 4) (Rodgers, 1967; Dean et al., 1988; Miller and Kent, 1988; Stamatakos et al., 1996; Spraggins and Dunne, 2001). Paleogeographic reconstructions of the Alleghanian orogeny must accommodate persistent late Paleozoic dextral displacement along the eastern Laurentian margin (Gates et al., 1988) followed by early Permian shortening in the southern Appalachians (Miller and Kent, 1988) and later northwest-vergent shortening in the central Appalachians (Stamatakos et al., 1996). Early Permian shortening in New England, determined on the basis of field and thermochronological data (Wintsch et al., 2003), may be coeval with shortening in the

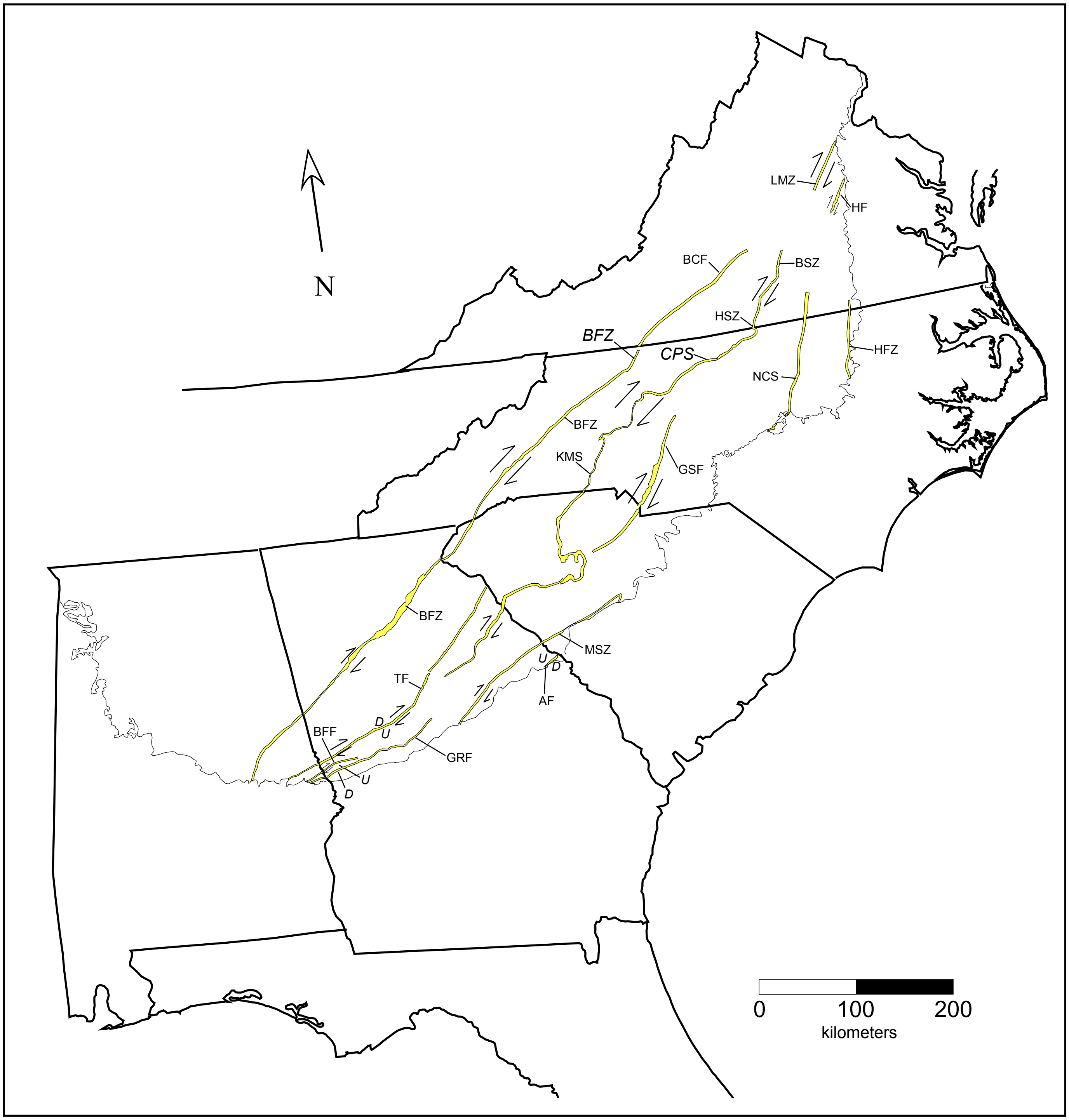

Figure 3.3. Distribution of major shear zones within the southern Appalachians, most of which exhibit dextral displacement. The fault zones are: LMZ-Lakeside mylonite zone, HF-Hylas fault, HFZ-Hollister fault zone, NCS-Nutbush Creek shear zone, BSZ-Brookneal shear zone (part of the CPS-Central Piedmont shear zone), BCF-Bowens Creek fault (part of the BFZ-Brevard fault zone), KMS-Kings Mountain shear zone (part of the CPS), GSF-Gold Hill-Silver Hill fault zone, MSZ-Modoc shear zone, AF-Augusta fault, TF-Towaliga fault (may be part of CPS), GRF-Goat Rock fault, and BFF-Bartletts Ferry fault. Fault distribution based on Hibbard (2000), Steltenpohl et al., (1992), Horton et al. (1989), and Gates et al. (1988). Some faults have significant normal displacement. The upthrown side is marked with a U, the downthrown side with a D.

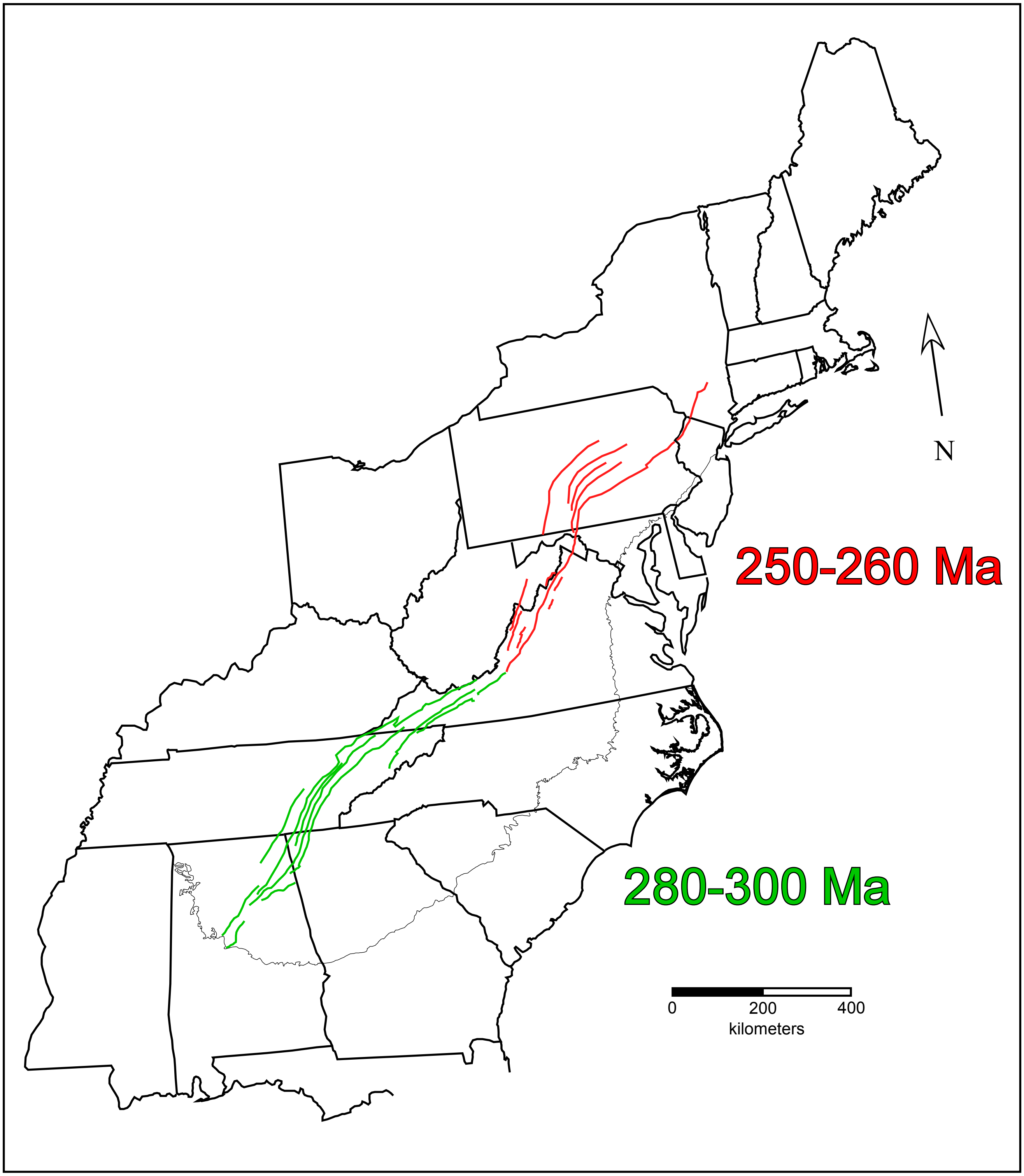

Figure 3.4. Map of Alleghanian foreland thrust belts in the Appalachians. The green lines show faults assigned to the southern Appalachian thrust belt, which propagated into the foreland earlier than did the central Appalachian thrust belt, represented by the red lines (Rodgers, 1967; Dean et al., 1988; Miller and Kent, 1988; Stamatakos et al., 1996). Ages were determined by matching magnetic inclination and declination of the secondarily remagnetized beds to Apparent Polar Wander Path ages (Miller and Kent, 1988; Stamatakos et al., 1996).

southern Appalachians. Rotational collision between Gondwana and Laurentia seems to be the most favorable approach to accommodate the complex kinematic history of the foreland. Shelley and Bossiere (2000) proposed clockwise rotation of Laurentia into Gondwana, which is generally consistent with the orientation of dextral shear zones and the temporal sequence of deformation (e.g., Dean et al, 1986; Miller and Kent, 1988; Spraggins and Dunne, 2002). Hatcher (2002) suggested that Gondwana first collided in New England, and rotated clockwise into the Laurentian margin, generating the central and southern Appalachian thrust belt through time. One serious drawback to this hypothesis is the implication that shortening in the central Appalachians predated that in the southern Appalachians.

Although the kinematic history of plate motion associated Alleghanian orogeny is complex, the sedimentary record of the event preserved in the Appalachian basin has been used as a case study by sedimentary petrologists for provenance analysis of foreland basins in collisional settings (e.g. Dickinson et al., 1983). Dickinson et al. (1983) suggested that the predominance of quartz in the Pennsylvanian sedimentary deposits reflects recycling of sedimentary deposits linked to the emplacement of thrust sheets. Other petrographic studies of sandstone framework grain composition of the early-middle Pennsylvanian Pocahontas, New River, and Kanahwa Formations in West Virginia interpret the sedimentary record to reflect unroofing of a batholithic sedimentary source related to the Alleghanian orogeny (Davis and Ehrlich, 1974; O'Connor, 1988). These interpretations were made without any attempt to integrate the complex kinematic history preserved in the Alleghanian orogen, which suggests a multi-stage tectonic assembly. As a result, the conclusions to be drawn from the petrographic studies of the Pennsylvanian-

Permian deposits in the Appalachian basin have the potential to contaminate studies of other foreland basins because of the improper inference of the Laurentian tectonic environment during the late Paleozoic. In order to interpret the late Paleozoic sedimentary record in the Appalachian basin, it is critical to develop a chronological tie between the style of Laurentian plate deformation and sedimentary deposition.

Study of the sedimentary record through time by use of K-Ar, $^{40}Ar/^{39}Ar$, and U-Pb dating of detrital minerals in the Appalachian foreland provides an additional objective method that can potentially help to resolve the kinematic history of the Alleghanian orogen. The K-Ar (and equivalent $^{40}Ar/^{39}Ar$) age of potassic minerals provides the age since which the mineral cooled below a characteristic closure temperature. The closure temperature is defined as the temperature at which diffusional loss of radiogenic daughter product from a mineral crystallographic lattice is statistically insignificant (Dodson, 1973). For example, the K/Ar age of muscovite records the time since the mineral cooled below 350 +/- 50 ^{o}C (McDougall and Harrison, 1999). Thermochronologic dates can be related to the exhumational history of a sedimentary source if the mineral examined is from a sedimentary deposit. U-Pb ages of zircons provide the age since the mineral was crystallized from a magma or experienced granulite-grade (>900 ^{o}C) metamorphism. U-Pb ages of detrital zircons can therefore be used to identify the source of sedimentary deposits by linking the ages to the genesis of possible source terranes. Because of the ever-increasing body of geochronological characterization of the Appalachian internides, radiometric sedimentary provenance applications provide a reliable link to the Alleghanian hinterland. The depositional record in the foreland can thus be compared to the tectonic history of the hinterland by more precise means.

In transpressional orogens, many of the tectonic features (such as accretionary prisms, volcanic arcs, and ophiolites) commonly associated with a collisional orogen and associated subduction zone may be lacking. In these tectonic environments, the collisional history along a plate margin may be more difficult to unravel by study of the sedimentary record. Because crustal shortening was an important part of the Alleghanian orogeny, and the periphery of the Laurentian margin is composed of a diverse mixture of terranes (e.g. Horton et al., 1989), it may be possible to see evidence of a progressive change in the age of the sedimentary source through time. Among these terranes, several are interpreted to be of Gondwanan origins (Fig. 5) and to have been accreted to Laurentia prior to the Alleghanian orogeny (e.g. Rast et al., 1976; O'Hara and Gromet, 1985; Hatcher et al., 1989; Hibbard, 2000). As a result, any early collisional episode between Gondwana and Laurentia would have involved crust of Gondwanan affinity. Gondwanan crust is distinct in age (Fig. 6) from that composing Laurentia by its involvement in the Late Proterozoic Pan-African/Brasiliano orogenies. In addition, several cratons in western Gondwana were created in the Paleoproterozoic (2000-2200 Ma) as part of the Transamazonian/Eburnian event (e.g. Cordani et al., 2000). By taking an inventory of detrital-zircon ages throughout the Pennsylvanian-Permian strata in the Appalachian basin, it may be possible to learn more about the complex tectonic history along the Laurentian margin during the late Paleozoic. To do this involves determining the relative contribution of detritus from Gondwanan terranes, and to determine when, or

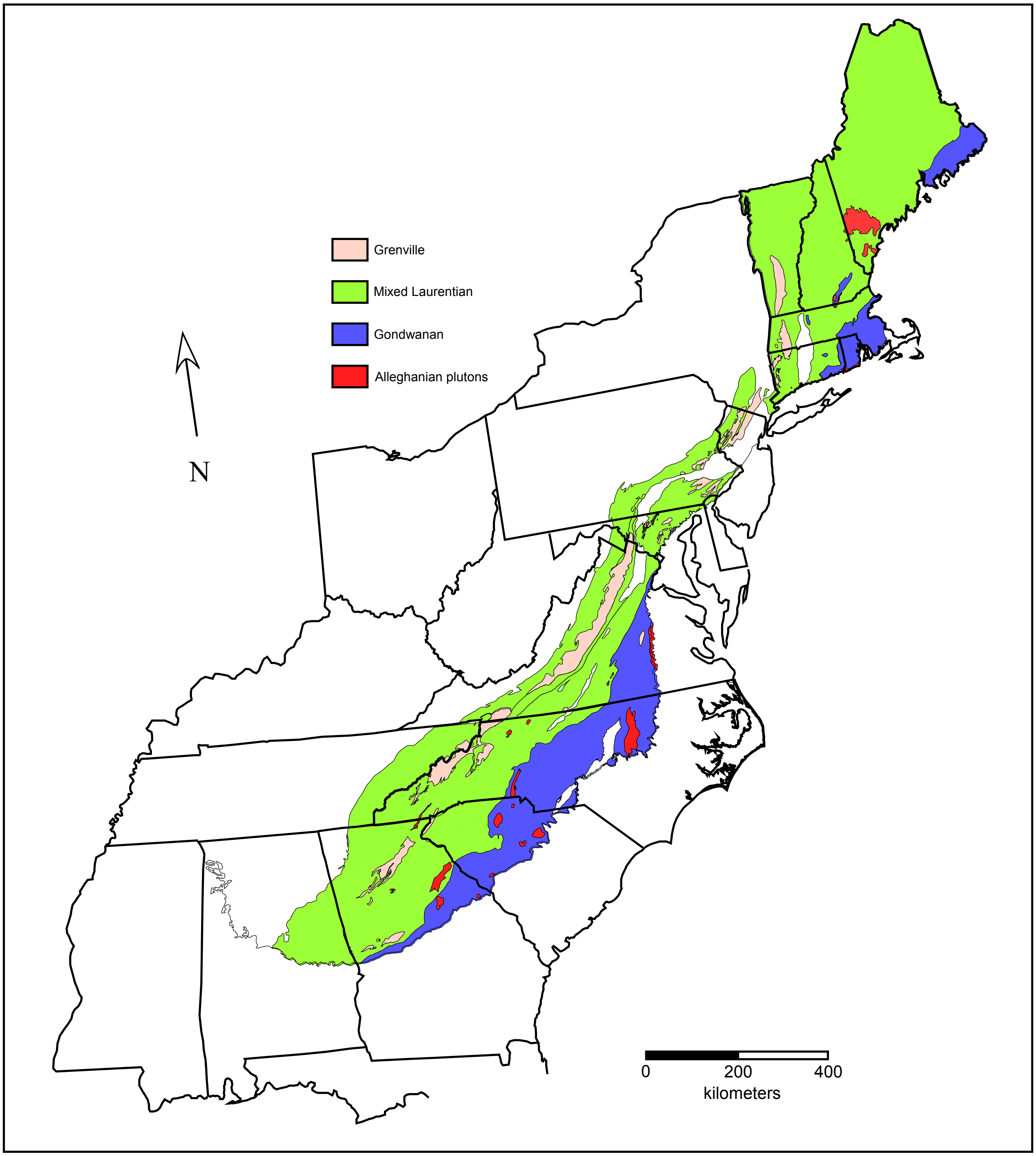

Figure 3.5. Distribution of Gondwanan and Laurentian crust in the Appalachian internides. Terrane boundaries are from Horton et al. (1989), Hibbard (2000), and Dorais et al. (2001). These terranes were probably accreted to Laurentia prior to the Alleghanian orogeny in the late Paleozoic. Ages that correspond to the various crustal provinces are displayed in Figure 3.6.

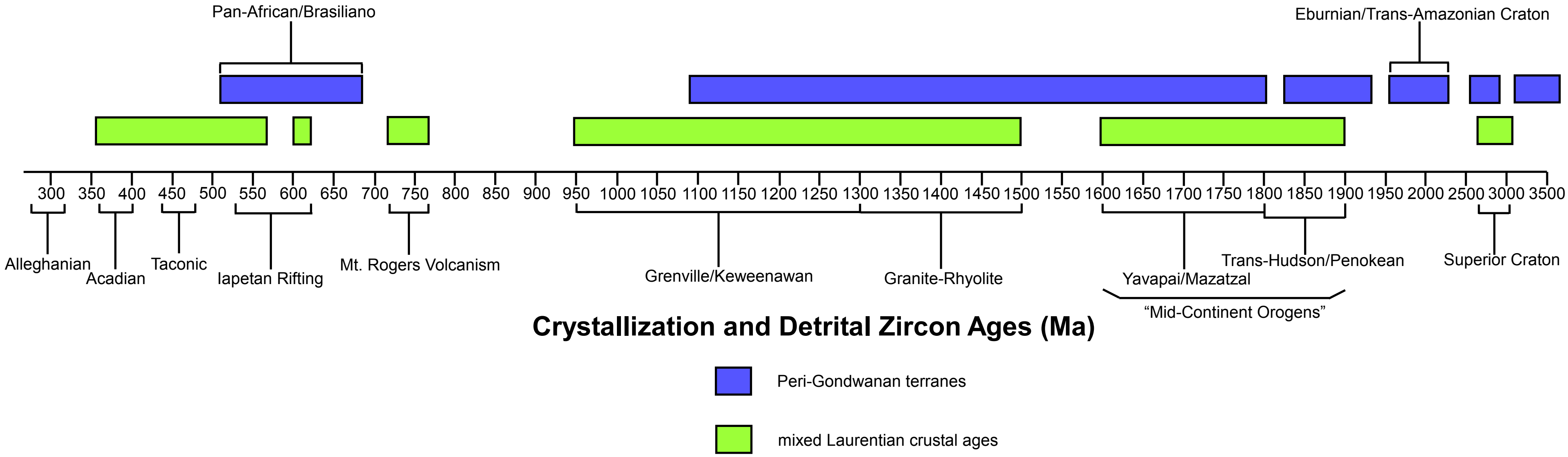

Figure 3.6. Bar graph representing composite zircon ages from peri-Gondwanan terranes (Samson et al., 2001; Coler and Samson, 2000; Wortman et al., 2000; Ingle-Jenkins et al., 1998; Mueller et al., 1994) and Laurentian crust (Hoffman, 1989; van Schmus et al., 1993; Aleinikoff et al., 1995).

if, Gondwanan crust was overthrust onto the Laurentian margin during the compressive phases of Alleghanian orogenesis.

Late Paleozoic Clastic Deposits in the Appalachian Basin

Pennsylvanian

The Pennsylvanian deposits in the Appalachian basin (Fig. 7) are recognized from an upward transition from Mississippian marine shales and reddish, hematite-rich sandstones and siltstones to coal-bearing, quartz-rich deposits. The Pennsylvanian can generally be categorized into three clastic wedges that distributed sediment semi-radially from orogenic sources along the eastern margin of Laurentia in the Pennsylvanian (Thomas, 1977; Chapter 2). These clastic wedges are the Mauch Chunk-Pottsville clastic wedge in Pennsylvania and Maryland; the Pennington-Lee clastic wedge in West Virginia, Virginia, Tennessee, Georgia, and Alabama; and the Floyd-Pottsville clastic wedge in central Alabama (Chapter 2).

The sedimentary record preserved within the Appalachian basin temporally overlaps with Alleghanian deformation in the Appalachian internides. Specifically, the Pennsylvanian strata were deposited during a well-documented phase of early Pennsylvanian dextral shear within the southern Appalachian hinterland (Gates et al., 1986; Stockey and Sutter, 1991; Wortman et al., 1998). Two recent studies (Thomas et al., 2004a; Chapter 2) used detrital-zircon-age analysis from the lower and middle Pennsylvanian deposits of the Pennington-Lee clastic wedge of Thomas (1977) to determine if textural and mineralogical characteristics of the Pennsylvanian deposits are related to a progressive shift in the source region in the context of orogenic exhumation

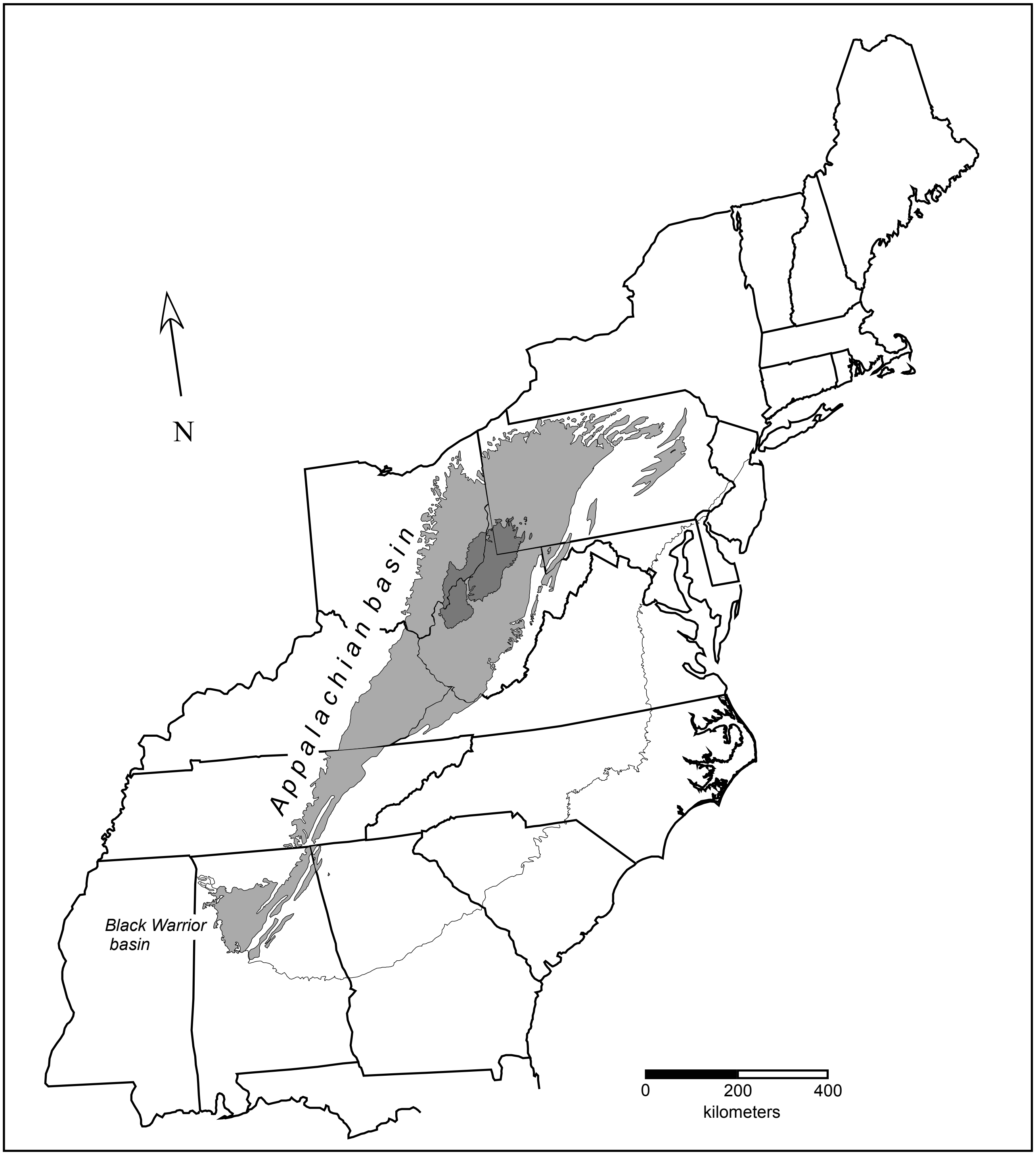

Figure 3.7. Outcrop area of Pennsylvanian and Permian deposits of the Appalachian and Black Warrior basins. Light gray shading corresponds to sedimentary deposits of Pennsylvanian age. Dark gray shading represents the Permian Dunkard Group.

of Gondwanan crust. The results indicate that the sediment is dominantly of Laurentian character, and that there is no substantial change in the population of detrital-zircon cores through the early to middle Pennsylvanian (Thomas et al., 2004a; Chapter 2).

Northwest-vergent folds of Alleghanian-age metamorphic isotherms and the southern Appalachian thrust belt are interpreted to have formed in the early Permian between 300-280 Ma (Secor et al., 1986; Miller and Kent, 1988; Adams and Su, 1996). The evidence of late Pennsylvanian/early Permian-age deformation comes from fold tests of an episodic, late Paleozoic remagnetization event in the Appalachian basin (Miller and Kent, 1988), translation and folding of Alleghanian-age metamorphic isotherms determined from $^{40}Ar/^{39}Ar$ (Secor et al., 1986), and $^{87}Rb/^{87}Sr$ ages of mylonites in the western Blue Ridge (Adams and Su, 1996). Several dextral shear zones, however, continued to be active throughout the formation of the west-vergent, collisional structures (Smits et al., 1988; Goldberg and Steltenpohl, 1990; Steltenpohl et al., 1992; Sacks and Secor, 1990; Kunk et al., 1995).

The youngest sedimentary deposits associated with the Alleghanian orogeny that are preserved in the Appalachian basin are the Permian Dunkard Group (Fig. 7). The Dunkard Group was likely deposited during the transition from dominantly dextral translation in the Pennsylvanian to early Permian northwest-vergent shortening within the southern Appalachian hinterland. Detrital-zircon ages from the Dunkard Group provide a test of the hypothesis that the source of sediment to the Appalachian basin changed during early Permian northwestward shortening as the collage of Gondwanan terranes was progressively displaced toward the craton.

K/Ar dating of detrital white mica also provides an opportunity to evaluate the sequence of unroofing in an orogenic belt through study of detritus in a foreland basin. Crustal thickening associated with plate collision drives prograde metamorphic reactions, and the synorogenic metamorphic rocks are progressively exhumed and eroded as the orogen continues to build. Examination of the detrital-white-mica (including muscovite and phengite) ages in the foreland basin reflect the cooling ages of the igneous and metamorphic rocks progressively exhumed in the orogenic belt. As the orogenic belt continues to evolve and to be exhumed from deeper crustal levels, the detrital-white-mica ages in successively younger strata in the foreland basin may trend toward values that nearly overlap with depositional age. Because the closure temperature of muscovite is estimated to be 350^{o} +/- 50^{o}C (McDougall and Harrison, 1999), exhumation of synorogenic upper greenschist/lower amphibolite facies metamorphic rocks during the orogenic event will yield synorogenic detrital-white-mica ages. In examples from modern orogenic belts, Willet and Brandon (2002) showed that thermochronometers have a tendency to reflect synorogenic cooling as the orogenic belt evolves toward a steady state (exhumation and erosion balanced by orogenic accretion). Within the Appalachians, Idleman (1996, 1998) has observed synorogenic ^{40}Ar/^{39}Ar ages in detrital muscovite of late Ordovician Taconic foreland sedimentary rocks in Newfoundland.

In a previous study, Aronson and Lewis (1994) analyzed detrital white mica from Devonian-Pennsylvanian strata in the Appalachian basin and noted a persistence of 371-419 Ma ages that corresponds to the Acadian orogeny, one of three major orogenic events that affected the Laurentian margin in the Paleozoic (see Chapter 1). Aronson and Lewis (1994) analyzed samples from the Upper Mississippian Mauch Chunk Formation, the

Lower Pennsylvanian Sharon Conglomerate, and the Middle Pennsylvanian Alleghany Formation in western Maryland and northeastern Ohio. The sample of Upper Mississippian Mauch Chunk Formation (collected in western Maryland) was considered part of the Mauch Chunk-Pottsville clastic wedge (Thomas, 1977). Stratigraphic and sedimentological studies of the Mauch Chunk in this region indicate that the sediment was supplied from a source in present eastern Pennsylvania (Hoque, 1968). The K/Ar age of a bulk sample of detrital white mica from the Mauch Chunk Formation is 377 +/- 10 Ma (Aronson and Lewis, 1994). By comparison, the lower Pennsylvanian Sharon Conglomerate and middle Pennsylvanian Allegheny Formation are both part of a south-southwesterly prograding fluvial system (Meckel, 1967; Edmunds et al., 1979, 1999). This fluvial system may have been part of a southwesterly prograding depositional system out of the New England region that was responsible for the deposition of the middle Pennsylvanian Sharp Mountain Member of the Pottsville Formation in eastern Pennsylvania (Robinson and Prave, 1995), and not associated with the Mauch Chunk-Pottsville clastic wedge of Thomas (1977). The detrital white mica K/Ar age is 371 +/- 10 Ma for the Sharon Conglomerate and 414 +/- 7 Ma for the Allegheny Formation (Aronson and Lewis, 1994). The shift to older ages may reflect mixing of older Taconian or Acadian metamorphic components from western New England (e.g. Harrison et al., 1989). It is significant that the ages do not reflect exhumation of Alleghanian syn-orogenic rocks, but imply a sedimentary source from rocks that were exhumed from 350 ^{o}C (~12 km depth) before the Alleghanian orogeny.

Permian

This study is focused on the Dunkard Group (Fig. 8) in southwestern Pennsylvania and eastern Ohio. The 360-meter-thick, late Paleozoic Dunkard Group, preserved within the northern Appalachian basin, was deposited in the latest Pennsylvanian or early Permian. Uncertainty surrounds the depositional age because of a lack of any well-defined marine taxa that can provide irrevocable proof of Permian age. On the basis of paleoflora and some terrestrial invertebrate remains, however, the Dunkard Group is considered to be early Permian age, and this is generally accepted (Martin, 1998; Edmunds, 1999).

The paleocurrent measurements, detrital grain size distributions, fluvial sandstone body trends, and interrelationships of sedimentary facies collectively indicate a paleoslope to the north-northwest for the central to northern Appalachian basin in southwestern Pennsylvania, southeastern Ohio, and northwestern West Virginia throughout the Pennsylvanian and Permian (Edmunds et al., 1979; Martin, 1998). The source of sediment to the basin is interpreted to be from the south or southeast. This would place the Dunkard Group at the northern edge of the much larger Pennington-Lee clastic wedge of Thomas (1977). The depositional environment of the Dunkard Group is interpreted to have shifted back and forth from north-northwestward prograding fluvial systems, to fluvial-swamp systems, and fluvial-lacustrine environments (Martin, 1998).

Sedimentologically, the Dunkard Group sandstones (Fig. 8) collectively plot in the "recycled orogenic" QFL field and the "quartzose recycled" QmFLt field of Dickinson et al. (1983). These results are identical to those from the Pennsylvanian sandstones of the Appalachian basin (Dickinson et al., 1983). Most sandstones in the upper Dunkard Group are very micaceous, and have subordinate sedimentary and

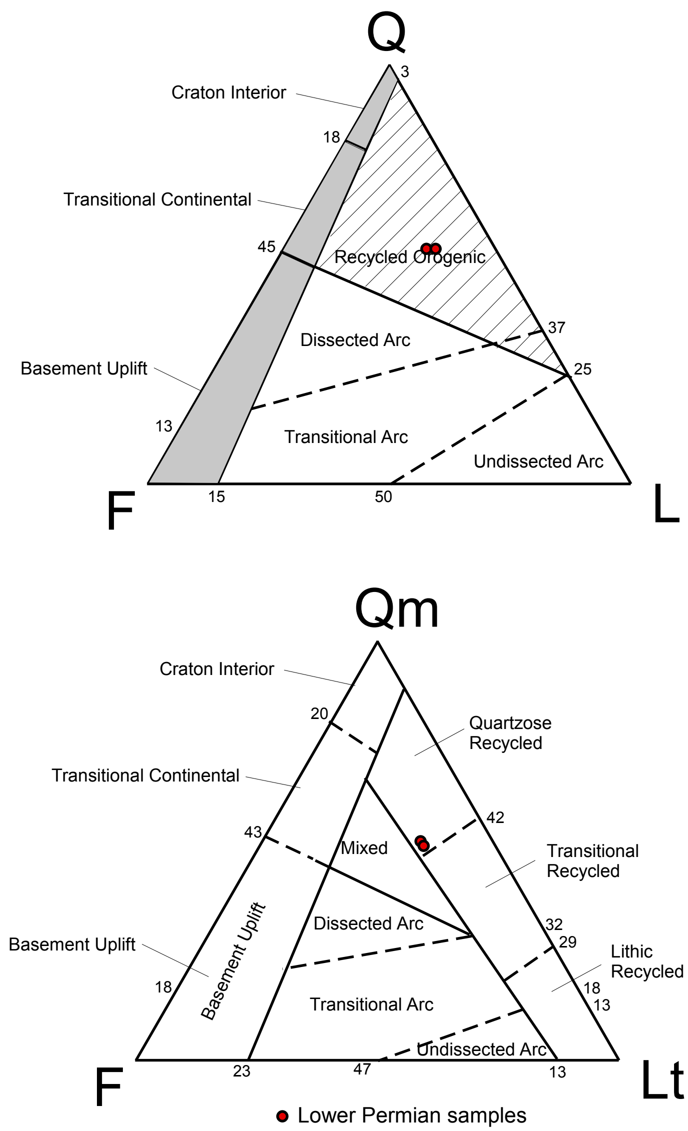

Figure 3.8. Plots of sandstone framework grain composition for various tectonic settings using the empirical relationships of Dickinson et al. (1983). The plotted samples are from the Permian sandstones of the Appalachian basin that were analyzed for detrital zircon ages. Q: all quartz (including chert), F: Feldspar (alkali and calcium), L: all lithic fragments, Qm: monocrystalline quartz, Lt: lithic fragments (including chert).

metasedimentary lithic fragments. Martin (1998) interprets the provenance of the Dunkard Group sandstones to be recycled rocks from the thrust belt of the southeastern Alleghanian orogen.

In Pennsylvania, the Dunkard Group is comprised of the Waynesburg, Washington, and Greene Formations (Edmunds, 1999). This nomenclature is difficult to apply throughout the Dunkard Group, however, because of the discontinuity of correlative marker beds. For example, the Waynesburg Formation is defined as the sandstones, shales, and minor coals between the Waynesburg coal at the base and the Washington coal at the top (Edmunds, 1999). The Waynesburg and Washington coals are absent in southeastern Ohio, and therefore the Waynesburg Formation cannot be distinguished from the overlying Washington Formation. The Waynesburg Formation is characterized essentially by an elongate north-south trending sandstone body (Martin and Henniger, 1969), and is generally interpreted to be part of a fluvial/deltaic plain (Martin, 1998). Martin and Henniger (1969) observed that this sandstone body is laterally discontinuous in all directions along the base of the Dunkard Group, and proposed that it is better identified as a "lentil", which they named the Mather lentil. On the basis of estimates of bankfull width for the Mather sandstone lentil of the Waynesburg Formation (Dominic, 1988), Martin (1998) estimated the length of the river system to be in excess of 300 km (185 miles). If correct, it would place the headwaters in the interior of the Alleghanian hinterland on the southeast. In this study, the Waynesburg Formation (of Edmunds, 1999) will be considered as undivided from the Washington Formation.

Sampling Locations

A sample of the Waynesburg sandstone, as defined by Martin and Henniger (1969), was collected along Interstate 79 near Waynesburg, Pennsylvania (Fig. 9). The stratigraphic context of the sample is given in the preceding paragraph.

The Upper Marietta sandstone (as referenced by Collins and Smith, 1977) is one of the more extensive beds in the Washington Formation in southeastern Ohio. It probably lies above the Waynesburg sandstone, but correlation is complicated by the discontinuous nature of the Waynesburg sandstone and any other chronostratigraphic marker bed into southeastern Ohio. The Upper Marietta typically is at the top of a sequence defined by the ~6-meter-thick Lower Marietta sandstone (as referenced by Collins and Smith, 1977) which is overlain by approximately 12 meters of distinctive maroon mudstone, called the Creston Reds (a paleosol) (as referenced by Collins and Smith, 1977). A sample of the Upper Marietta sandstone was collected and supplied by Dr. Greg Nadon of Ohio University near Athens, Ohio (Fig. 9).

Another sample from the Washington Formation was collected from Belpre, Ohio, across the Ohio River from Parkersburg, West Virginia (Fig. 9). The sampling location corresponds to a measured section published in Collins and Smith (1977, p. 40). The sample comes from a 3-foot-thick sandstone bed 44 meters (145 feet) above the Upper Marietta sandstone, which places it in near the top of the Washington Formation at about the stratigraphic level of the regionally correlative Hundred Sandstone (Martin, 1998). The sample is medium grained, micaceous, and somewhat friable. Petrographic analysis of framework grains reveals that the sample is 53% quartz, with 15% feldspar, and 32% lithic fragments of siltstone, chert, chloritic slate, phyllite, and schist. In the

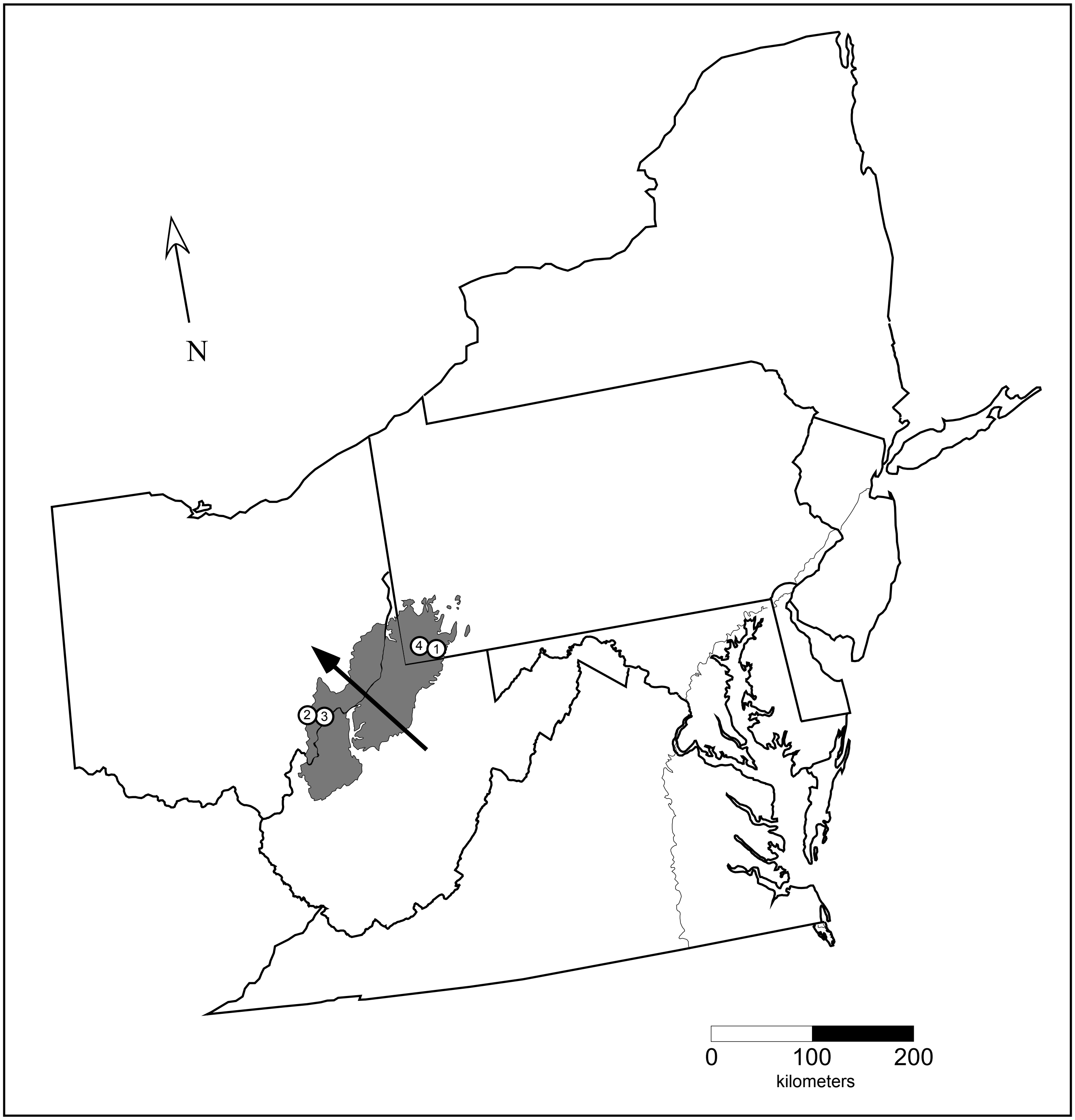

Figure 3.9. Sample locations from the Dunkard Group: 1- Waynesburg sandstone, Waynesburg, Pennsylvania; 2-Upper Marietta sandstone, Athens County, Ohio; 3-Washington/Greene Formation, Belpre, Ohio; 4-Greene Formation, Rush Crossroads, Pennsylvania. The arrow on the map is the mean paleocurrent direction for the Dunkard Group (N38ºW) (Martin, 1998).

tectonic discrimination diagram of Dickinson et al. (1983), the sample plots as “recycled orogenic” in QFL space and “quartzose recycled” in QmFLt space (Fig. 8, Table 1).

The Greene Formation is the stratigraphically highest in the Dunkard Group. The Upper Washington freshwater limestone defines the base (Edmunds, 1999). Similar to the Washington Formation, the Greene does not contain any marine beds. In southwestern Pennsylvania, the Greene Formation is >500 feet thick, and is composed of an interbedded sequence of shale, sandstone, freshwater limestones, and thin coals (Roen, 1972). The sample collected at Rush Crossroads, Pennsylvania in the Holbrook 7.5-minute quadrangle (Fig. 9) is approximately 220 feet above the top of the Washington Formation and 960 feet above the Pittsburgh coal seam, as determined by extrapolation of structural contours from Roen (1972). The sample comes from an approximately five-foot-thick sandstone bed that thins laterally to shale and fine-grained micaceous sandstone. It is medium to fine grained, and is also somewhat friable. Petrographically, the sample is nearly identical to the Washington Formation sandstone from Belpre, Ohio, and plots in the same tectonic environment in QFL space (Fig. 8; Table 1).

Methods

The Permian-age sandstones analyzed for detrital white mica were crushed on a clean steel plate. The disaggregated sand was sieved into size fractions of >250 microns, 180-250 microns, and 110-180 microns. The sand from each size fraction was reground on a clean porcelain block. Equant grains were powdered into a much finer fraction, and the entire aliquot was passed through a sieve to remove the pulverized fraction. Shaking the mica separate on clean paper allowed equant grains to roll off and segregated the white mica separate. The muscovite fraction was further purified by sonification in

	Petrologic constituents (%)					Particle	standard	
	Q	***Qm***	***F***	***L***	***Lt***	***size (phi)***	***deviation (phi)***	***skewness***
L. Pennsylvanian sandstones								
Tumbling Run Member, eastern PA	92	92	0	8	8	-3.5	0.79	-0.05
Pottsville Formation, central PA	97	97	2	1	1	-3.48	0.63	-0.09
Pocahontas Formation, southern WV	71	68	9	20	23	1.87	0.63	0.05
Lee Formation, western VA	94	93	3	3	4	2.23	0.72	0.32
Raccoon Mountain Formation, northeastern GA	97	96	1	3	4	2.01	0.57	0.11
Montevallo Coal Zone, central AL	48	33	5	47	62	-3.3	1.08	-0.08
L. Permian sandstones								
Washington Formation, southeastern OH	56	53	15	28	32.5	2.63	0.57	0.05
Greene Formation, southwestern PA	53	51	15	32	34	3.18	0.45	0.00

Table 3.1. Petrologic characteristics of Pennsylvanian and Permian sandstone samples from the Appalachian basin that were processed for detrital zircons. Constituents were classified using the Gazzi-Dickinson criteria (Qm: monocrystalline quartz; Q: Qm + chert; F: feldspar (alkali + calcium); L: lithic fragments (not including chert); Lt: lithic fragments (including chert). Petrographic classification of the sedimentary framework constituents was determined by point counts on >300 grains of thin sections stained for feldspars. Clasts were defined as any particle greater than 0.2 cm. Particle size, standard deviation, and skewness were determined using the techniques outlined in Blatt (1992).

ethanol to dislodge any lithic fragments or oxides. This separate was passed through a magnetic separator to enhance removal of phases that might have substantial alteration. Finally, the sample was handpicked to remove any remaining impure phases. The samples were then split for determination of K and radiogenic ^{40}Ar concentrations. Potassium concentrations were obtained by flame photometry on $LiBO_3$-fused sample splits under the supervision of Dr. James Aronson at Case Western Reserve University. The concentration of radiogenic ^{40}Ar was determined on the noble gas extraction line at Case Western Reserve University under the supervision of Dr. James Aronson. The analyses began by fusing the sample in an evacuated glass extraction line. The evolved gas was stripped of any non-noble elements by exposure to hot and cold titanium wool and an SAE getter, and analyzed on an MS-10 mass spectrometer. Calibration is done by analysis of LP-6 biotite standard, for which a concentration of 19.03×10^{-10} moles ^{40}Ar/gm is routinely obtained.

Two samples of the Permian Dunkard Group were collected for U-Pb age analysis of the detrital-zircon population. The two samples of Permian sandstone analyzed for detrital zircons were processed separately from the samples used for detrital white mica dating. Prior to crushing, the samples were washed to remove any potential contamination. After each sample was crushed, all components of the jaw crusher and disk mill were cleaned thoroughly to avoid any potential cross-contamination between samples. The resulting fine sand was separated into populations by density using a Wilfley table. The two densest fractions were collected and dried, and ferromagnetic phases were removed using a hand magnet. The remaining nonmagnetic dense mineral fraction was poured into a separatory funnel with doubly filtered pure 1, 1, 2, 2 tetrabromoethane (density =2.967 grams/milliliter). The dense split was collected, dried, and sieved into two size fractions. The 106-150 micron size fraction was hand-picked for detrital zircons, which were subsequently divided into populations based on color (yellow, pink, opaque, etc.) and morphology (faceted or round), similar to the procedure outlined in Thomas et al. (2004a) and in Chapter 2. Representative samples from each of the color/morphologic categories were included in each composite sample. The purpose of this approach was to try to identify and include all of the potential sources. The detrital-zircon populations were embedded in an epoxy mount, which was polished to expose the cores.

U-Pb geochronology of detrital-zircon cores was conducted with a Micromass Isoprobe ICPMS equipped with nine faraday collectors, an axial Daly detector, and four ion-counting channels. The Isoprobe is connected to a New Wave DUV 193 laser ablation system, which has an emission wavelength of 193 nm. The analyses were conducted on 35-micron spots with an output energy of ~32 mJ and a repetition rate of 8 hz. Each analysis consisted of one 20-second integration on backgrounds (on peak centers with no laser firing) and twenty 1-second integrations on peaks with the laser firing. The depth of each ablation pit is ~20 microns. The collector configuration allows simultaneous measurement of ^{204}Pb in a secondary electron multiplier while ^{206}Pb, ^{207}Pb, ^{208}Pb, ^{232}Th, and ^{238}U are measured with Faraday detectors. All analyses were conducted in static mode.

Inter-element fractionation during the analysis was monitored by analyzing fragments of a large concordant zircon crystal that has a known (ID-TIMS) age of 564 ± 4 Ma (G.E. Gehrels, unpublished data). This reference zircon was analyzed once for every four unknowns. The calibration correction contributes ~3% (2-sigma) systematic uncertainty to both ^{206}Pb / ^{238}U and $^{206}Pb/^{207}Pb$ ages. The isotopic ratios are also corrected for common Pb using the measured ^{204}Pb, assuming an initial Pb composition according to Stacey and Kramers (1975) and uncertainties of 1.0 and 0.3, respectively, for $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$. The reported ages are based primarily on $^{206}Pb/^{207}Pb$ ratios for >1200 Ma grains and $^{206}Pb/^{238}U$ ratios for <1200 Ma grains (Appendix B).

During data reduction, the $^{238}U/^{206}Pb$ ratio was analyzed for consistency throughout the 20 one-second peak integrations. Ideally, the $^{238}U/^{206}Pb$ ratio remains constant (+/- 15%). A few grains with relatively large uncertainties displayed two well-defined plateaux connected by a few ratios of intermediate value to the plateaux. This pattern is interpreted to represent penetration of a growth zone by the laser ablation process during the course of the analysis. For these grains, each plateau was reduced as an individual age, so that both core and rim could be assessed. For the purpose of provenance analysis, we are interested in determining the origin of the sediment from a magmatic source. Therefore, the youngest age obtained for the zircons with complex growth histories was taken to represent the age of the source terrane. Detrital zircons with complex growth histories are annotated in Table 2. Several plutons in the Appalachians were crystallized in a zirconium-saturated melt so that xenocrystic cores are commonly preserved within the younger zircons (Watson and Harrison, 1983), making identification of Pennsylvanian-age zircons difficult (Zartman and Hermes, 1987; Rice et al, 1994; Whitehead and Gromet, 1997; Miller et al., 2000; Miller et al., in review). Detrital-zircon ages are supposed to be representative of the crust which produced sediment. In cases where an intruded magma is supersaturated with zirconium, an inherited zircon that is incorporated into the melt may not be completely resorbed (Watson and Harrison, 1983), so the age of the crust forming event may not be represented by analysis of the inherited core. These cases may be resolved by analyzing the U/Th ratios of the rim overgrowth. U/Th ratios <10 are indicative of growth in a magma rather than growth during a high-grade metamorphic event (G. Gehrels, pers. comm.) which may not form new crustal sources of sediment.

Results

Washington Formation

K/Ar analyses

Detrital white mica K/Ar ages from the Waynesburg sandstone are 392.7 +/- 9.2 Ma for the 250-180 micron size fraction and 394.9 +/- 9.2 Ma for the 180-110 micron size fraction (Table 2). A sample of >250 micron white mica from the Upper Marietta sandstone has a K/Ar age of 391.8 +/- 9.2. The stratigraphically highest sample, the Washington/Greene Formation of Belpre, Ohio, has a K/Ar age of 394 +/- 9.3 Ma for a >250 micron bulk sample of detrital white mica. A 180-250 micron size white mica fraction from the same sample has an age of 390.8 +/- 9.2 Ma. All of these samples suggest a sedimentary source that cooled below ~350^{o}C in the early Devonian.

U-Pb analysis

The stratigraphically lowest sample of the two samples from the Dunkard Group is the one described from Belpre, Ohio. It is near the top of the Washington Formation (labeled Washington Formation in Fig. 10). U-Pb ages of detrital zircons (Fig. 10, Appendix B) have modes at 315-440 Ma and 900-1350 Ma, and several smaller peaks at 540-565 Ma, 650-780 Ma, 1440-1500 Ma, and 1650-1680 Ma (Appendix B). U/Th ratios

Sample	**Size fraction (microns)**	**$1x10^{-10}$ moles $^{40*}Ar$**	**K_2O (wt. %)**	**Age (Ma)**
Greene/Washington Fm. (OH)	>250	60.05	9.476	394.0 + 9.3
	180-250	64.491	10.27	390.8 + 9.2
Marietta Sandstone (Washington Fm.; OH)	>250	66.063	10.49	391.8 + 9.2
Waynesburg Sandstone (Washington Fm.; PA)	180-250	65.53	10.38	392.7 + 9.2
	110-180	64.141	10.094	394.9 + 9.2

Table 3.2. Detrital white mica K/Ar analytical data from the Permian Dunkard Group sandstones in the Appalachian basin.

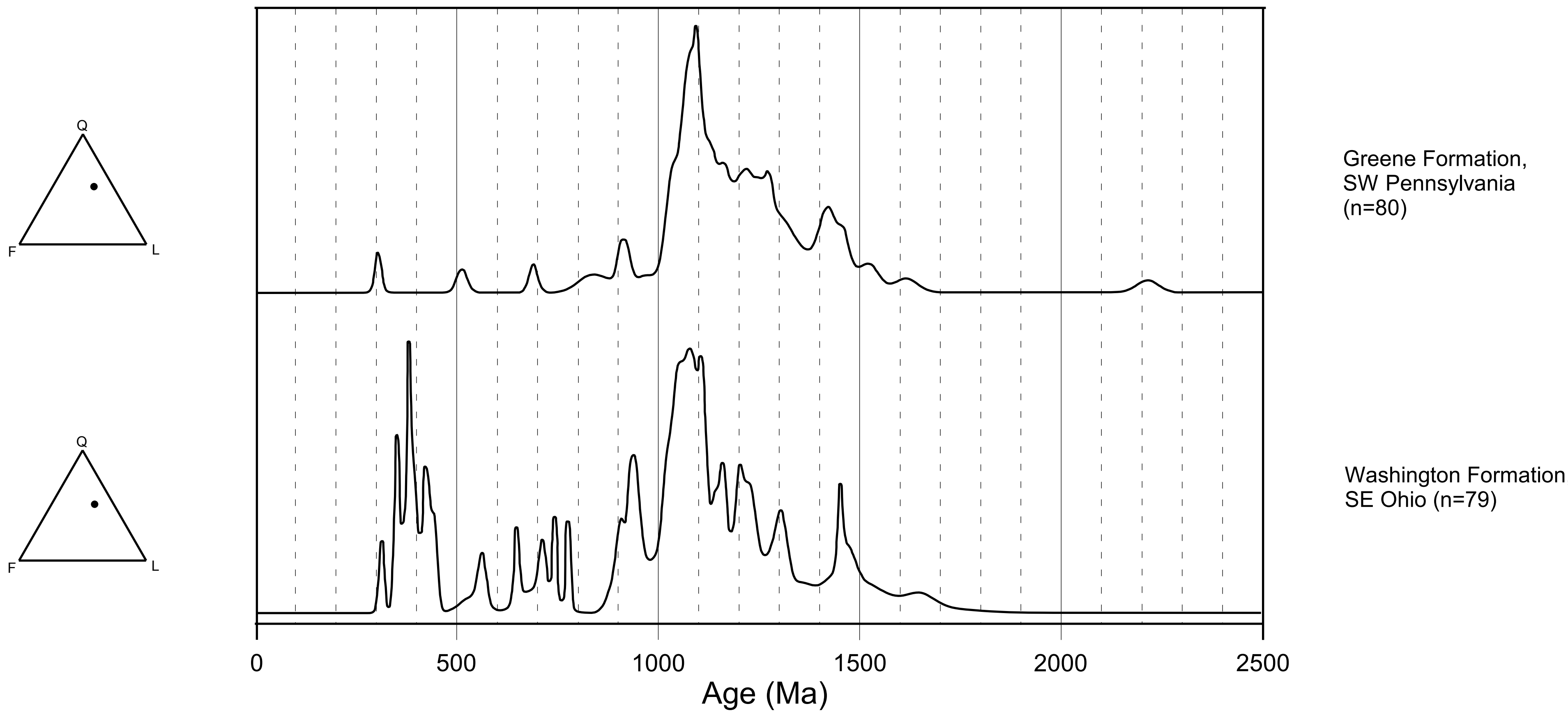

Figure 3.10. Frequency-probability plot for detrital zircons from the early Permian Washington Formation and overlying Greene Formation. The Washington Formation has a much larger proportion of Paleozoic-age zircons. Both of the Permian samples have modes at about 1100 Ma.

of the analyzed zircons are <6 (Appendix B), implying that the zircons are of igneous derivation. One detrital zircon with a complex growth history has an upper plateau age of 1108.4 +/- 128.4 Ma and a distinct lower plateau age of 314.5 +/- 13.1 Ma with a U/Th ratio of 1.0, suggesting that it formed in a zirconium-saturated melt around an older xenocrystic core. The 314.5 +/- 13.1 Ma age represents the first evidence that synorogenically derived rocks from the Alleghanian orogeny were incorporated into drainages that emptied into the Appalachian basin.

The Washington Formation sample has a significant population (~25%) of acicular, pink zircon grains. Although no clear relationship between age and morphology can be demonstrated, it is notable that several of those zircons have middle Paleozoic (360-440 Ma) ages (Appendix B).

Greene Formation

U-Pb analysis

The detrital-zircon population of the sample from the Greene Formation is dominated by a large mode at 1000-1300 Ma, and secondary modes at 900-960 Ma, and 1380-1500 Ma (Fig. 10; Appendix B). Single zircon ages at 305.4 +/- 15 Ma, 513.3 +/- 26.2 Ma, 691.4 +/- 21.9 Ma, 824 +/- 61.9 Ma, 1529.2 +/- 40.5 Ma, 1620.7 +/- 51.5 Ma, and 2219+/- 52.9 Ma define the rest of the detrital zircon age spectra. U/Th ratios of the analyzed zircons are all <6 (Appendix B), implying an igneous origin. In contrast to the Washington Formation sample, zircons with acicular morphology are absent from the Greene Formation sample, as are the middle-Paleozoic-age zircons.

Discussion

K/Ar ages of detrital white mica from the Dunkard Group are very similar to ages obtained from Devonian-Pennsylvanian age deposits by Aronson and Lewis (1994). One serious drawback to dating populations of detrital white mica by the K/Ar method is the inability to resolve individual age components. For example, if the detrital white mica is comprised of a mix of Ordovician- and Pennsylvanian-age populations, the resulting Devonian K/Ar age may be geologically meaningless. These concerns are somewhat abated by single-grain $^{40}Ar/^{39}Ar$ analyses on white mica from the Pennsylvanian-age Llewellen Formation of eastern Pennsylvania, in which the micas are all of Devonian age (B.D. Idleman, pers. communication). In addition, there is no significant variation between the K/Ar ages of different size fractions from the same sample. If multiple sources of different age contributed white mica to the Dunkard Group detritus, the ages may be fractioned by size. The lack of variability suggests that the mica-bearing source had been uniformly exhumed in the early Devonian. Therefore, the source rocks for sediment comprising the Dunkard Group cooled below ~350^{o}C during the early Devonian, implying that the sedimentary source resided at shallow levels within the crust and was not significantly affected by Alleghanian metamorphism. The consistency of Devonian K/Ar ages of detrital white mica in the late Paleozoic strata of the Appalachian basin indicates that the Alleghanian orogenic belt must have been maintained by significant lateral translation, not vertical uplift. Analogous relationships between translational kinematic history and thermochronologic ages have been documented in Taiwan (Willet et al., 2003). Therefore, it is suggested that the Devonian ages are reflective of the transpressional evolution of the Alleghanian orogen throughout the Pennsylvanian.

The detrital-zircon-age populations in the Dunkard Group samples have gross similarity to those reported from other studies of Pennsylvanian age deposits (Gray and Zeitler, 1997; Thomas et al., 2004a; Eriksson et al., 2004; Chapter 2). Figure 11 is a comparison of compiled detrital zircon ages from Pennsylvanian deposits (Thomas et al., 2004a; Eriksson et al., 2004; Chapter 2) with Permian deposits (this study) in the Appalachian basin from a source within the central and southern Appalachian orogen. Both the Pennsylvanian- and Permian-age deposits have detrital-zircon populations dominated by zircons of Grenville (900-1300 Ma) age. Both deposits also have distinct early-middle Paleozoic age detrital-zircon-age populations. In detail, some potentially significant differences characterize the Permian detrital-zircon-age population from those of the Pennsylvanian, such as the lack of any Archean grains, and ages that overlap with the Central Plains orogens. The Permian deposits also have a more pronounced peak (650-800 Ma) of Neoproterozoic-age grains, and a few grains of Pennsylvanian age.

Although Neoproterozoic detrital zircon ages (650-800 Ma) are most commonly attributed to Gondwana (e.g. Mueller et al., 1994; Thomas et al., 2004a,b), possibly some of these detrital zircons could be of Laurentian origin. Lukert and Banks (1984) reported a zircon U-Pb age of 732 +/- 5 Ma for the Robertson River pluton, north of Charlottesville, in central Virginia. The Robertson River pluton may be related to ~750 Ma volcanism at Mount Rogers (Aleinikoff et al., 1995), the ~740 Ma Crossnore Complex (Su et al., 1994), and the 734 +/- 26 Ma Bakersville granite/gabbro intrusive suite in North Carolina (Goldberg et al., 1986) that are collectively interpreted to represent a failed rifting event in the Neoproterozoic (Aleinikoff et al., 1995; Tollo and Hutson, 1996). Field relationships show that most of these igneous rocks of very limited

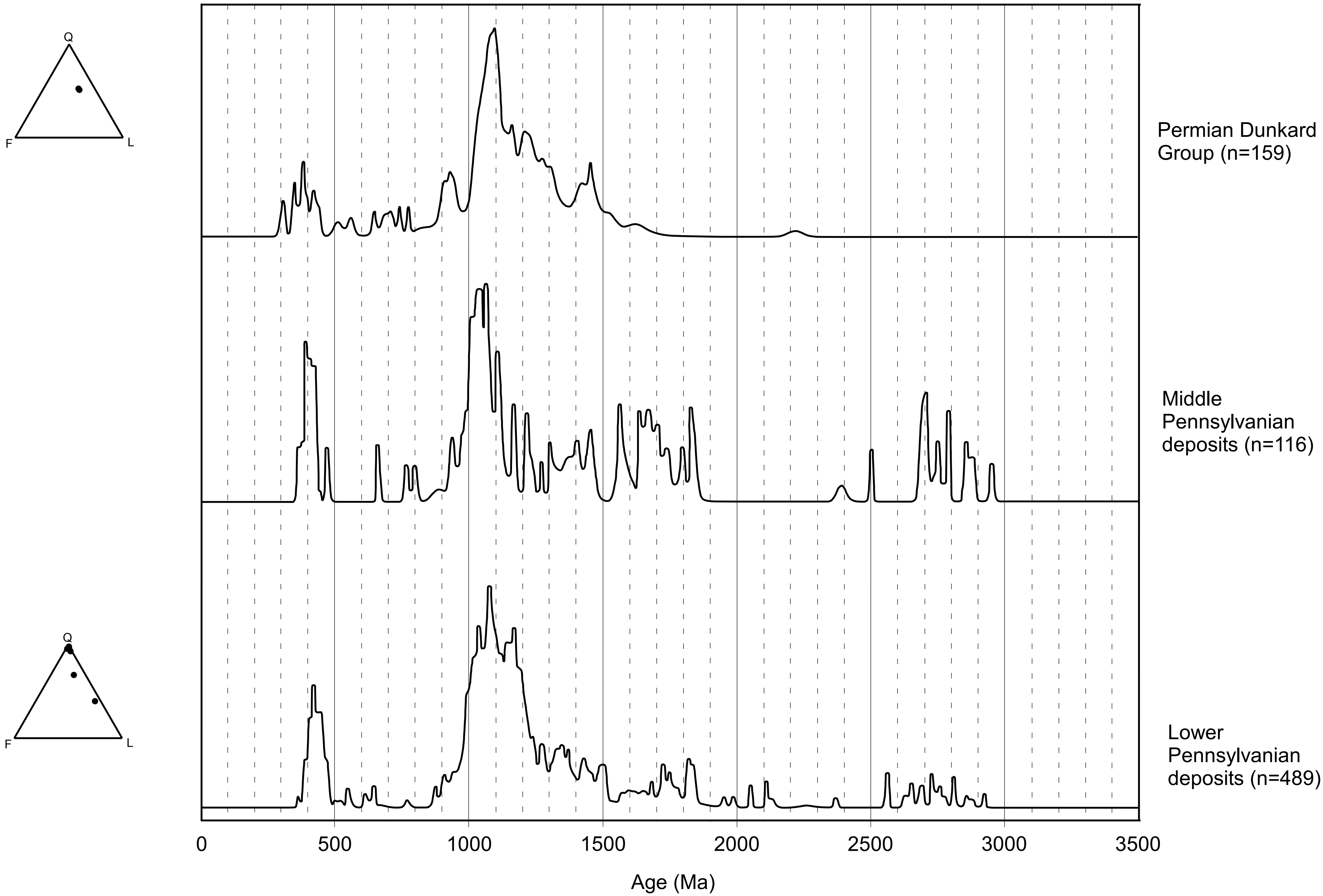

Figure 3.11. Composite frequency plot of detrital zircon ages from Pennsylvanian and Permian sandstones from the Appalachian basin. Lower Pennsylvanian sandstones include the Pottsville Formation, Pocahontas Formation, Lee Formation, lower Raleigh Sandstone, Sewanee Conglomerate, and Raccoon Mountain Formation (from Thomas et al., 2004; Eriksson et al., 2004; Chapter 2-this dissertation). The middle Pennsylvanian detrital-zircon-age population includes the upper Raleigh sandstone (Eriksson et al., 2004) and Cross Mountain Formation (Thomas et al., 2004). The Permian detrital-zircon-age data are from this study.

exposure and extent are unconformably overlain by lower Paleozoic strata, so that there is no reason to believe that these Neoproterozoic volcanic rocks were a substantial source of detritus into the Appalachian basin. However, it is also possible that detritus from these Neoproterozoic sources was incorporated in the lower Cambrian synrift and passive margin strata, and subsequently recycled (e.g. Thomas et al., 2004a). Detrital-zircon populations from the late Proterozoic Ocoee Supergroup, a late Proterozoic rift basin fill (Rast and Kohles, 1986) that overlies the Neoproterozoic plutons, do not contain Neoproterozoic (650-800 Ma) zircons (Bream, 2002). The lack of 650-800 Ma zircons in the eastern Laurentian pre-orogenic strata suggests a greater likelihood that late Proterozoic detrital zircons are from Gondwanan sources, and not recycled from Laurentian cover.

Thomas et al. (2004a) interpret the presence of 1600-1950 Ma and Archean (>2500 Ma) detrital zircon grains in the Pennsylvanian-age deposits to reflect recycling of late Precambrian-early Cambrian synrift and Cambrian-Ordovician passive margin strata during the early stages of the Alleghanian orogeny. In the composite frequency plot of lower Pennsylvanian detrital zircon ages (Fig. 11), the 1600-1950 Ma and Archean grains comprise 12.7% of the 488 analyzed grains. Middle Pennsylvanian sandstones (Cross Mountain Formation and Upper Raleigh sandstone of the New River Formation) from the Appalachian basin have a much higher percentage of 1600-1950 Ma and Archean-age detrital zircons (26% of 104 analyzed). The lack of Paleoproterozoic and Archean detrital-zircon-age populations in the Permian sandstones suggests the lower Paleozoic cover sequence over the orogenic internides had been completely removed by Permian time. These ages are not common in the zircon population of the Blue Ridge or

Inner Piedmont, on the basis of results obtained by Bream (2002) and Bream et al. (2004).

Petrographically, most of the basal Pennsylvanian sandstones analyzed for detrital zircon ages (Chapter 2) plot in the same QFL tectonic environment as the Permian sandstones (Fig. 12). However, the detrital-zircon-age populations are markedly different, as is the interpreted style of deformation along the Laurentian margin. The Pennsylvanian sandstones have lower proportions of feldspar than the Permian deposits (Table 1; Fig. 12). The slightly higher concentration of feldspar may accompany a significant change in the tectonic environment that cannot be detected using the QFL tectonic discrimination diagram of Dickinson et al. (1983).

Interpretation

Using Martin's (1998) Dunkard Group average paleocurrent direction (N38^{o}W), the Dunkard Group sediment was likely derived from a southern Appalachian source (Fig. 9). Temporal control on the tectonic history of the southern Appalachian hinterland can be constrained by $^{40}Ar/^{39}Ar$ thermochronology on the crystalline basement, timing of motion on major dextral shear zones in the region, and estimates of the timing of thrusting into the Appalachian foreland.

Available geochronological constraints on the timing of dextral shear in this region suggest that this phase of deformation was active relatively early in the orogenic event. The Brevard shear zone, a 750-km-long shear zone separating the Laurentian Inner Piedmont from the eastern Blue Ridge province (Bream, 2002), is the most conspicuous fault in the southern Appalachians. It exhibits pervasive dextral shear that

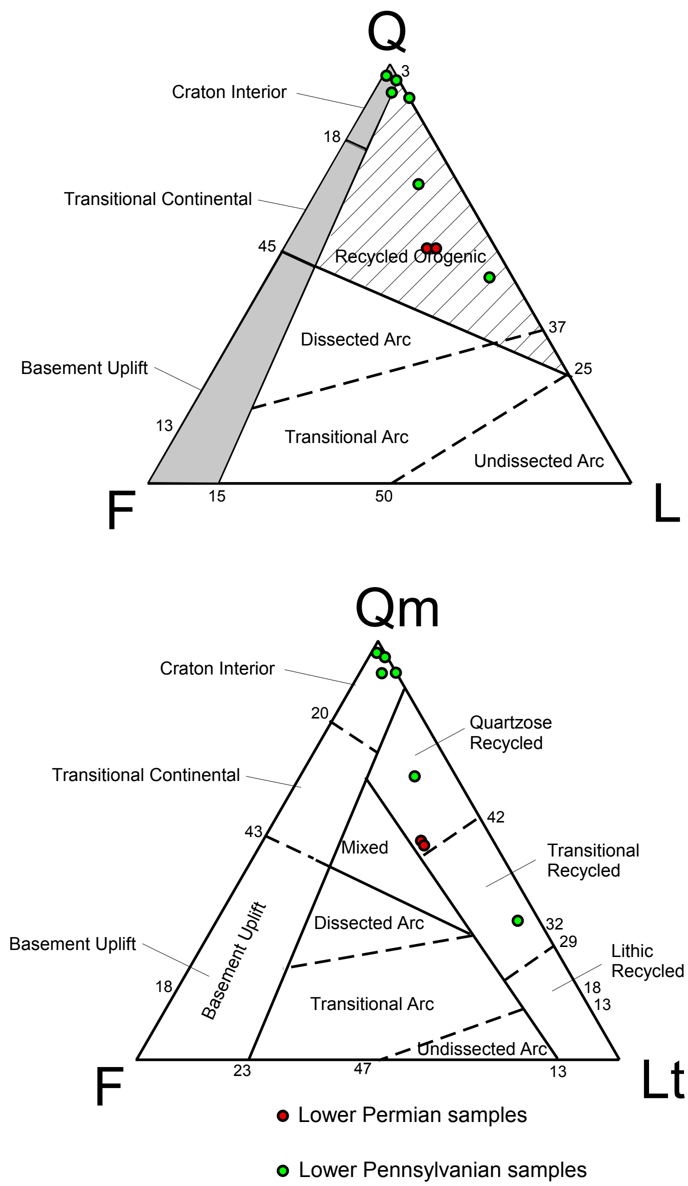

Figure 3.12. Plots of sandstone framework grain composition for various tectonic settings using the empirical relationships of Dickinson et al. (1983). The plotted samples are from the Pennsylvanian and Permian sandstones of the Appalachian basin that were analyzed for detrital-zircon ages. Q: all quartz (including chert), F: Feldspar (alkali and calcium), L: all lithic fragments, Qm: monocrystalline quartz, Lt: lithic fragments (including chert).

was later overprinted by brittle, top-to-the-north thrusting (Hatcher, 2001). It is constrained to have been active at about 315 Ma (Stockey and Sutter, 1991), on the basis of $^{40}Ar/^{39}Ar$ ages of hornblende and white mica. The 300-324 Ma $^{87}Rb/^{87}Sr$ age of a cross-cutting, synkinematic dike along the Brookneal shear zone in southern Virginia, considered the northern trace of the Central Piedmont shear zone (Hibbard, 2000), constrains the timing of movement (Fig. 3) (Gates et al., 1988). The Hyco shear zone is the southern equivalent of the Brookneal shear zone, and the timing of displacement is constrained at 327-319 Ma, on the basis of high-precision U-Pb dates of syn-kinematic granitic dikes that crosscut the shear zone (Wortman et al., 1998). There is no published estimate of the timing of dextral motion on the Hylas fault (Fig. 3), which separates the Goochland and Raleigh terranes, but the 330 +/- 8 Ma Petersburg granite (Wright et al., 1975) exhibits dextral shear deformation (Fig. 3) (Gates et al., 1988). Although deformation associated with dextral shear was likely an important factor in the development of topography during the deposition of the late Mississippian-early Pennsylvanian deposits, it probably was not as important in Permian time.

West-vergent shortening in the southern Appalachians was likely to have been the dominant deformational process during the early Permian. Secor et al. (1986) compiled $^{40}Ar/^{39}Ar$ data in the southern Appalachians in South Carolina to resolve the timing of folding of metamorphic isotherms. In their review, they suggested that northwest-vergent folding occurred from 280 to 300 Ma. $^{87}Rb/^{87}Sr$ ages of west-vergent mylonites in the Blue Ridge of northwestern North Carolina yield ages of about 300 Ma (Adams and Su, 1996). Timing of folding in the southern Appalachian thrust belt, that relies on fold tests of episodically remagnetized strata, also suggests west-vergent shortening at about 280-

300 Ma (Miller and Kent, 1988), which predates deformation in the central Appalachians (Stamatakos et al., 1996). Because the southern Appalachian thrust belt extends from central Alabama to the Virginia recess in southern Virginia (Rodgers, 1967; Dean et al., 1988), the proposed sediment source for the Permian beds was likely involved in this deformational event.

A compilation of $^{40}Ar/^{39}Ar$ thermochronologic data for hornblende and white mica from this region (Figs. 13 and 14, Appendix C) indicates that the southeastern Virginia cooled rapidly from prograde Alleghanian metamorphism in the Late Pennsylvania-early Permian. Rapid cooling is linked to rapid exhumation, so that these areas may represent potential sources of sediment to the Dunkard Group (Fig. 14). The K/Ar ages from detrital white mica are Devonian, indicating that the source of mica, if not a mixture of various age populations, is from a relatively shallow crustal source. It is therefore likely that the mica was associated with Taconic and Acadian metasedimentary rocks in the region of the Blue Ridge physiographic province, or perhaps the Carolina terrane or Goochland terrane. A metasedimentary source is further supported by sedimentary petrography of the Dunkard Group sediments (Martin, 1998).

The composition of the igneous and metamorphic Blue Ridge physiographic province is dominated by Grenville-age crust (Bream, 2002; Bream et al., 2004). The eastern Blue Ridge also has a complex metamorphic history, which is variably attributed to the Taconic or Acadian orogenies (e.g., Hibbard, 2000; Hatcher, 2001). It is possible that the upward transition in the detrital-zircon-age populations in the Pennsylvanian and Permian deposits in the Appalachian basin reflects uplift of the Blue Ridge during the Alleghanian orogeny from relatively shallow levels of the crust.

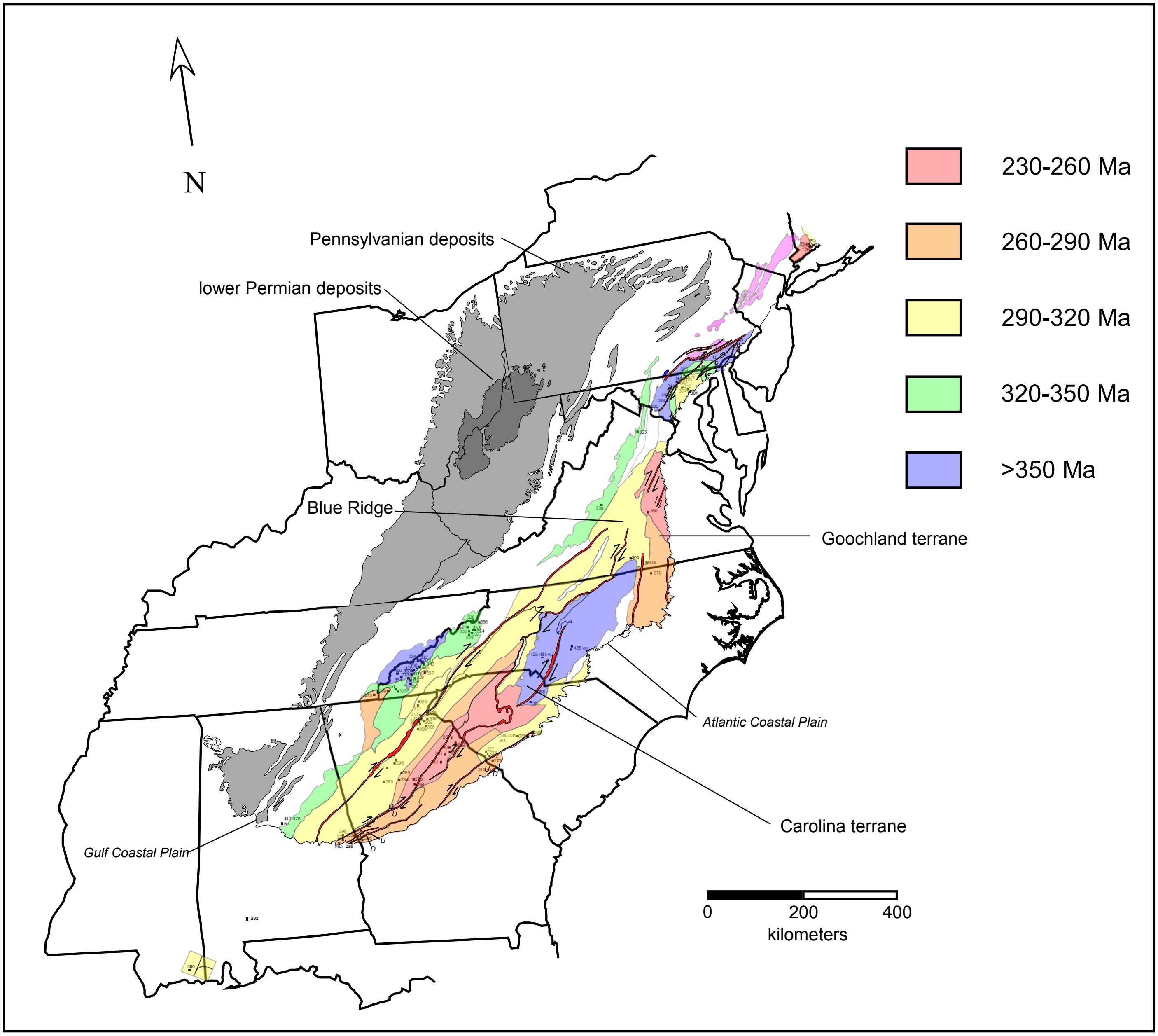

Figure 3.13. Compilation of K/Ar and $^{40}Ar/^{39}Ar$ ages of muscovite from the Appalachian crystalline hinterland. The age reflects the time that the rock cooled below the closure temperature for muscovite, which is approximately 350° (McDougall and Harrison, 1999). Despite being in the core of the orogen, a substantial proportion of the Carolina terrane did not experience significant exhumation (>15 km) during the Alleghanian orogeny. The Dunkard Group sediment must have been supplied from regions northeast or northwest of the northern Carolina terrane (Blue Ridge or Goochland provinces). Bold red lines mark the location of dextral shear zones in the Alleghanian orogen. References are listed in Appendix A.

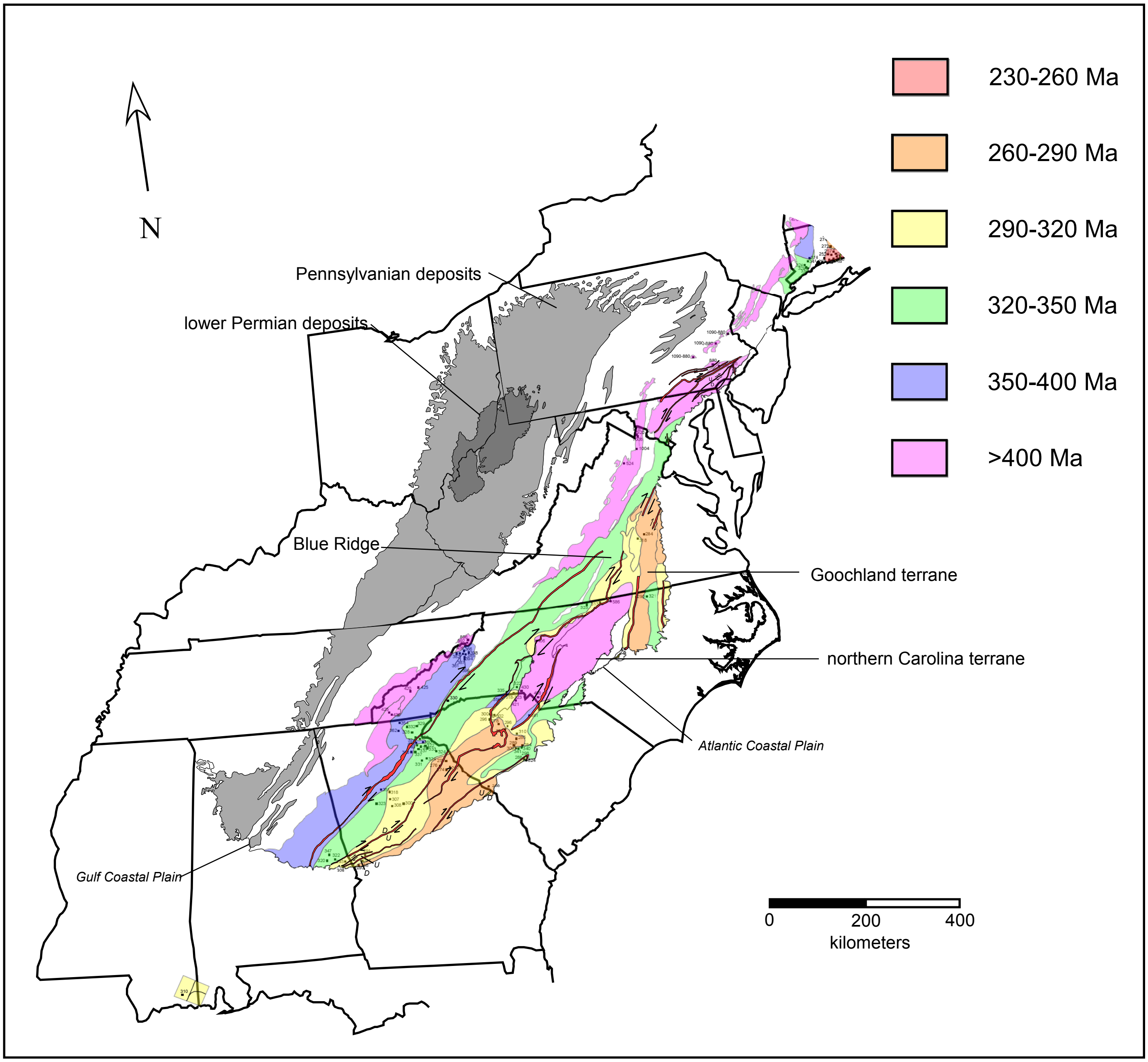

Figure 3.14. Compilation of K/Ar and $^{40}Ar/^{39}Ar$ ages of hornblende from the Appalachian crystalline hinterland. The age reflects the time that the rock cooled below the closure temperature for hornblende, which is approximately 500^{o} (McDougall and Harrison, 1999). Because hornblende is not as sensitive to argon loss as muscovite, Alleghanian (<320 Ma) ages highlight regions that experienced significant syn-tectonic amphibolite facies metamorphism, and not merely exhumation. The syn-Alleghanian metamorphic ages in the region surrounding the Goochland terrane suggests that this region was significantly uplifted during Alleghanian deformation, although the deformation likely post-dated deposition of the Dunkard Group sediment. Bold red lines mark the location of dextral shear zones in the Alleghanian orogen. References are listed in Appendix B.

The Goochland terrane, comprised predominantly of Grenville-age crust, is another possible source of sediment for the Dunkard Group. The Goochland terrane is positioned east of the Milton and Chopawamsic terranes, which are interpreted to represent Ordovician arcs that were accreted during the Taconic orogeny (Coler et al. 2000). To the south, along strike of the Goochland terrane, lies the unequivocally Gondwanan Carolina terrane (Secor et al., 1983). Recent geochronological studies of the Goochland terrane in central and southern Virginia revealed several plutons of Neoproterozoic age (654-588 Ma) that intrude Grenville age (~1100 Ma) basement (Owens and Tucker, 2003). The Neoproterozoic plutons temporally overlap with the Pan-African/Brasiliano orogenies that are commonly found in Gondwana. The Goochland terrane is, therefore, likely to represent an accreted terrane of Gondwanan origin that might have provided sediment to the Appalachian basin in the early Permian.

Ordovician-age zircons in the detrital-zircon population of the Washington Formation may signal incorporation of the Milton/Chopawamsic terrane as the Laurentian margin was translated toward the Appalachian basin. The lack of Ordovician-age detrital zircons in the Greene Formation may simply reflect a variation in the age of the crust underlying the different drainage networks. The lack of Archean-age detrital zircons, and presence of one Paleoproterozoic-age (2200 Ma) and Neoproterozoic (580-680 Ma) populations that overlap with Eburnian/Transamazonian and Brasiliano/Pan-African age events in Gondwana suggest that a non-Laurentian source terrane may have been integrated into the drainage network.

Detrital-zircon ages from modern rivers headed in the Alleghanian igneous and metamorphic hinterland may help to identify what age populations are characteristic of these regions. Eriksson et al (2003) published detrital-zircon ages from several major rivers presently draining the Appalachian region, including the Susquehanna, Potomac, James, Savannah New, and Ohio Rivers. Of these, only the Savannah River drainage is completely isolated within the Alleghanian crystalline hinterland. The James River drainage extends into regions underlain by lower Paleozoic strata in the Appalachian thrust belt of western Virginia. Detrital-zircon-age populations from the Susquehanna and Potomac Rivers are not considered because they drain regions underlain by Pennsylvanian-age deposits in the Appalachian basin. Denudation rates along the Atlantic slope of the Piedmont (much of it underlain by peri-Gondwanan terranes) are much lower (4.5-7 m/m.y.; Pavich et al., 1985; Pavich, 1986) than in the higher-standing Blue Ridge (~30 m/m.y. Matmon et al., 2003) and folded Paleozoic strata (~27 m/m.y.; Sevon, 1989). Therefore, it is expected that detrital-zircon ages reflect a sedimentary source from the high-standing crystalline Blue Ridge and lower Paleozoic strata. This hypothesis is strengthened by detrital-zircon ages from metasedimentary rocks of the Blue Ridge (Bream, 2002), which are similar to the detrital-zircon-age populations from the James and Savannah Rivers (Eriksson et al., 2003). The detrital-zircon-age populations from the James and Savannah Rivers are very similar to those from the Permian Dunkard Group revealing a dominance of Mesoproterozoic (900-1300 Ma) ages and a secondary mode of early to middle Paleozoic ages (400-550 Ma) (Fig. 15) (Eriksson et al., 2003). The modern high-standing Blue Ridge is comprised chiefly of weakly metamorphosed lower Paleozoic strata, Precambrian synrift sedimentary and igneous rocks, and Grenville basement, which might have been compositionally similar to the westward-vergent thrust sheets exhumed in the early Permian of the Alleghanian

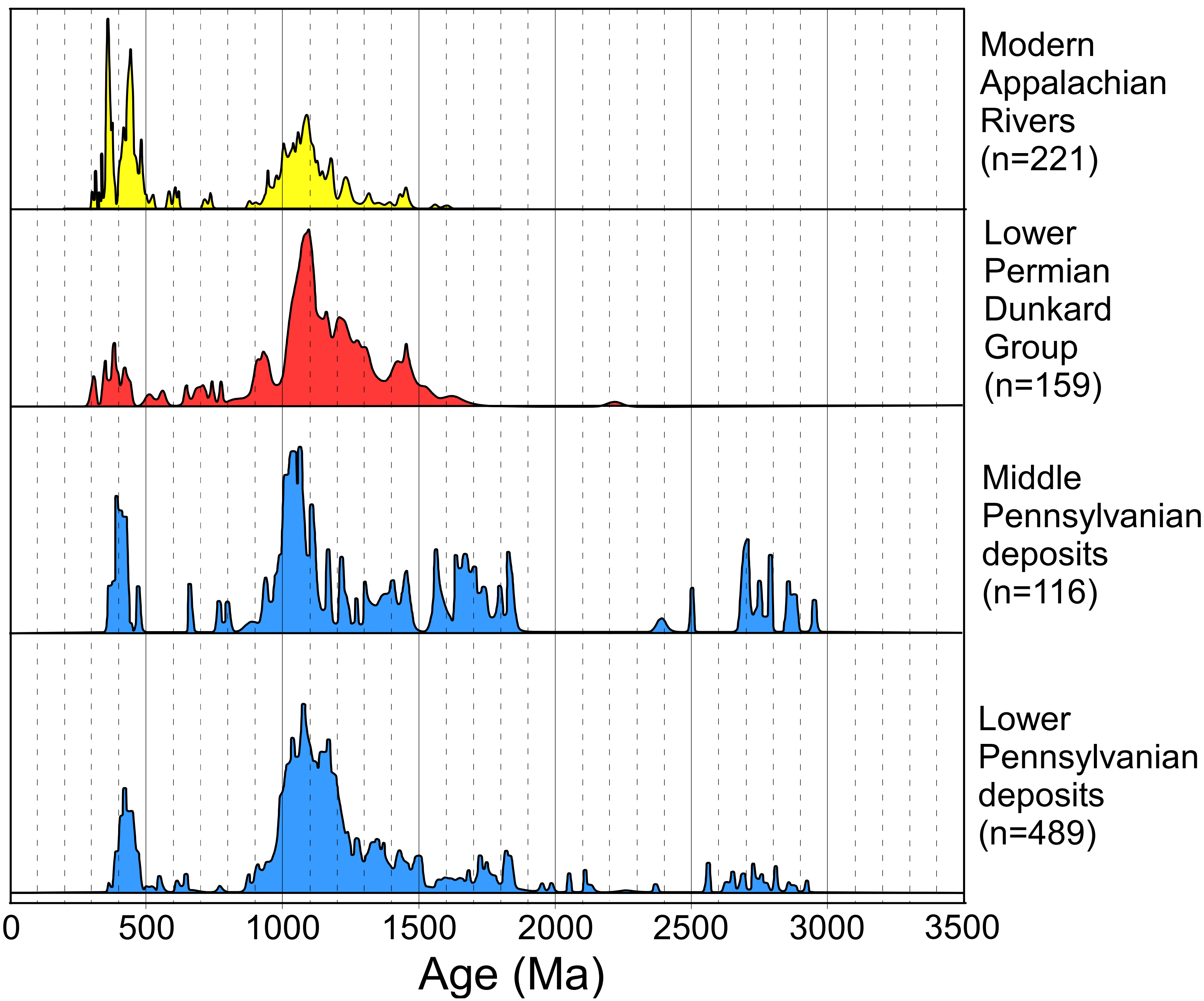

Figure 3.15 Detrital-zircon-age populations from the James and Savannah Rivers compare favorably with those from the lower Permian Dunkard Group. The enhanced Paleozoic mode in the modern river detrital-zircon ages is attributable to the Savannah River zircons. The Savannah River drains the Inner Piedmont, which is intruded by numerous late Paleozoic plutons (Samson, 2001).

orogeny. Therefore, the highest-standing modern vestiges of the Alleghanian orogenic hinterland may have been compositionally similar to that which supplied sediment to the lower Permian sandstones.

Conclusion

A subtle, but persistent, shift in the detrital-zircon-age population in the Permian may reflect a change from dominantly transpressional deformation in the Pennsylvanian to northwestward shortening associated with collision in the Permian. The changes are not dramatic, and may reflect the oblique nature of the collisional orogen. Transpressional orogens may not have features that are diagnostic of continental subduction-margin collisions, such as a volcanic arc or accretionary prism.

Shallow levels of the crust continued to be the dominant source of detritus to the northern Appalachian basin throughout the early Permian. The implication of this result is that the tectonic history of the Alleghanian collision is difficult to ascertain by examination of the sedimentary record preserved in the Appalachian basin. Traditional petrographic approaches (e.g. Dickinson et al., 1983) cannot resolve some of the potentially important changes in source regions that could be linked to significant tectonic events in the source region. Models of the tectonic history of the Alleghanian orogeny reliant on petrography of sedimentary deposits are not sufficient to characterize the history of deformation in the hinterland.

Petrographic and radiometric sedimentary provenance proxies indicate that the early Permian Alleghanian highlands were dominated by Laurentian crust that was metamorphosed or cooled during the Devonian. This is consistent with a source in the Blue Ridge (Bream, 2002; Bream et al., 2004), although a source from the peri-

Gondwanan Goochland terrane or Carolina terrane cannot be eliminated. Crustal contributions from peri-Gondwanan terranes did not dominate the sedimentary load delivered to the Appalachian basin in the late Paleozoic.

CHAPTER 4: PALEOENVIRONMENTAL PARAMETERS OF THE LATE PALEOZOIC APPALACHIAN BASIN DEDUCED FROM STABLE ISOTOPIC ANALYSIS OF LACUSTRINE CARBONATES, SOUTHWESTERN PENNSYLVANIA

Introduction

The paleoenvironmental characteristics of the Pennsylvanian System represent a dramatic shift in global paleoecology, which may have ties to high atmospheric concentrations of oxygen (Fig. 1) (Berner, 2002), low concentrations of carbon dioxide (Fig. 1) (Mora et al., 1996; Retallack, 2001), changes in global atmospheric circulation (Parrish, 1993; Tabor and Montanez, 2002), and the growth of continental glaciers on Gondwana (Caputo and Crowell, 1985; Mii et al., 1999). Within the Appalachian basin, the Pennsylvanian marks a dramatic shift in the paleoenvironment from the Mississippian, despite the similarity in tectonic setting (Donaldson et al., 1985). The population of plant assemblages is interpreted to represent a shift from a drier climate in the Mississippian, to a wetter, more tropical environment in the early Pennsylvanian (Fig. 1) (Pfefferkorn and Thomson, 1982; Cecil, 1990; Valero Garces et al., 1997; Dimichele et al., 2001). In the late Pennsylvanian/early Permian, floral assemblages indicate a shift back to a warmer-drier climate (Pfefferkorn and Thomson, 1982; DiMichele et al., 2001). These climatic shifts within the Appalachian basin are likely linked to changes in latitude as eastern Laurentia drifted from the southern dry latitudes through the paleo-intertropical convergence zone, and into the northern dry latitudes (Cecil, 1990; Soreghan et al., 2002), and/or may also reflect changes in atmospheric circulation in response to the uplift of the Alleghanian hinterland (Parrish, 1993).

The Pennsylvanian-Permian stratigraphy in southwestern Pennsylvania represents the most complete and continuous record of sedimentation throughout the late Paleozoic

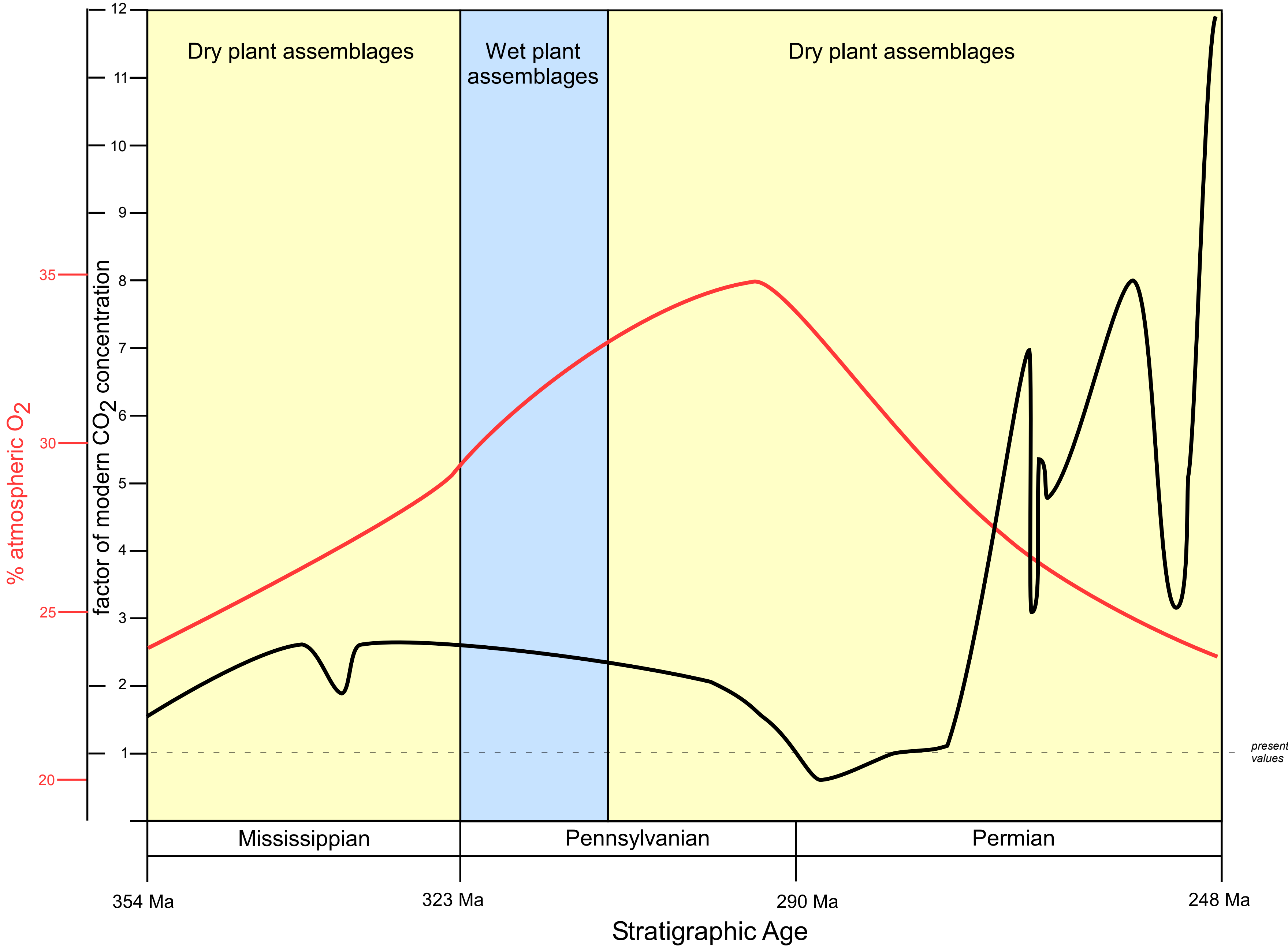

Figure 4.1. Global Mississippian-Permian paleoenvironmental conditions, including the relative concentration of atmospheric CO_2 (from Mora et al., 1996; Retallack, 2001) and O_2 (Berner et al., 2000, 2003). The colored regions correspond to paleoenvironment based on paleofloral assemblages (Pffefferkorn and Thompson, 1982; DiMichele et al., 2001).

in the Appalachian basin. The deposits are the erosional product of the coeval late Paleozoic Alleghanian orogeny along the eastern margin of Laurentia. Within southwestern Pennsylvania, the Pennsylvanian System is commonly broken into four major subdivisions (from oldest to youngest): the Pottsville Formation, the Allegheny Formation, the Conemaugh Group, and the Monongahela Group (Fig. 2) (Edmunds et al., 1999). The Permian Dunkard Group is comprised of the basal Washington and overlying Greene Formations (Martin, 1998).

Within the Pennsylvanian and Permian strata in the northern Appalachian basin, non-fossiliferous limestone is commonly intercalated with terrestrial coal deposits and paleosols. The oldest limestone deposits are middle-late Desmoinian, and their significance is attributed to factors such as tectonism and changes in global paleoclimate (Weedman, 1989; Cecil, 1990), although these interpretations are controversial (e.g., Valero-Garces et al., 1997). The limestones were presumably deposited in a system of freshwater lakes in interdistributary depressions (Williams, 1968; Weedman, 1989). The appearance of freshwater limestones in the Appalachian basin coincides with paleobotanical evidence of climate change. Near the Westphalian-Stephanian European stage boundary (Late Desmoinian, North American stage), there was a dramatic shift in the population of land plants from wetlands in the Early to Middle Pennsylvanian to varieties adapted to more arid conditions (Fig. 1) (Pfefferkorn and Thompson, 1982; DiMichele et al., 2001). Similarly, the depositionally oldest freshwater limestone from the Pennsylvanian-age deposits in southwestern Pennsylvania is the Johnstown limestone in the early Desmoinian Allegheny Formation (Fig. 3)(Weedman, 1989; Edmunds et al., 1999). Is this merely a coincidence, or did these paleolakes form in response to regional

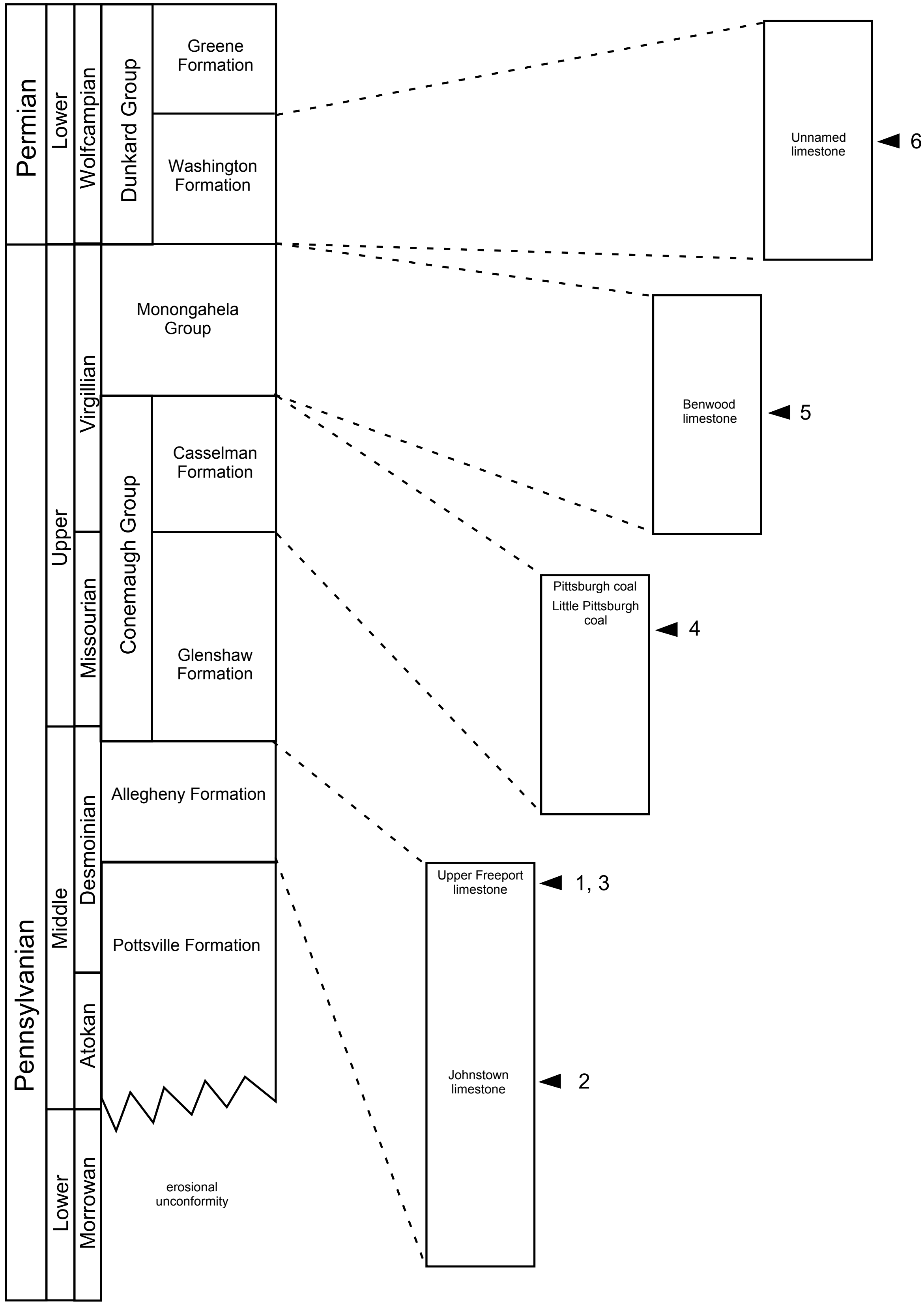

Figure 4.2. Pennsylvanian-Permian stratigraphy of southwestern Pennsylvania (after Edmunds et al., 1999). Numbers in the stratigraphic column refer to the stratigraphic position of sample locations in Figure 4.3.

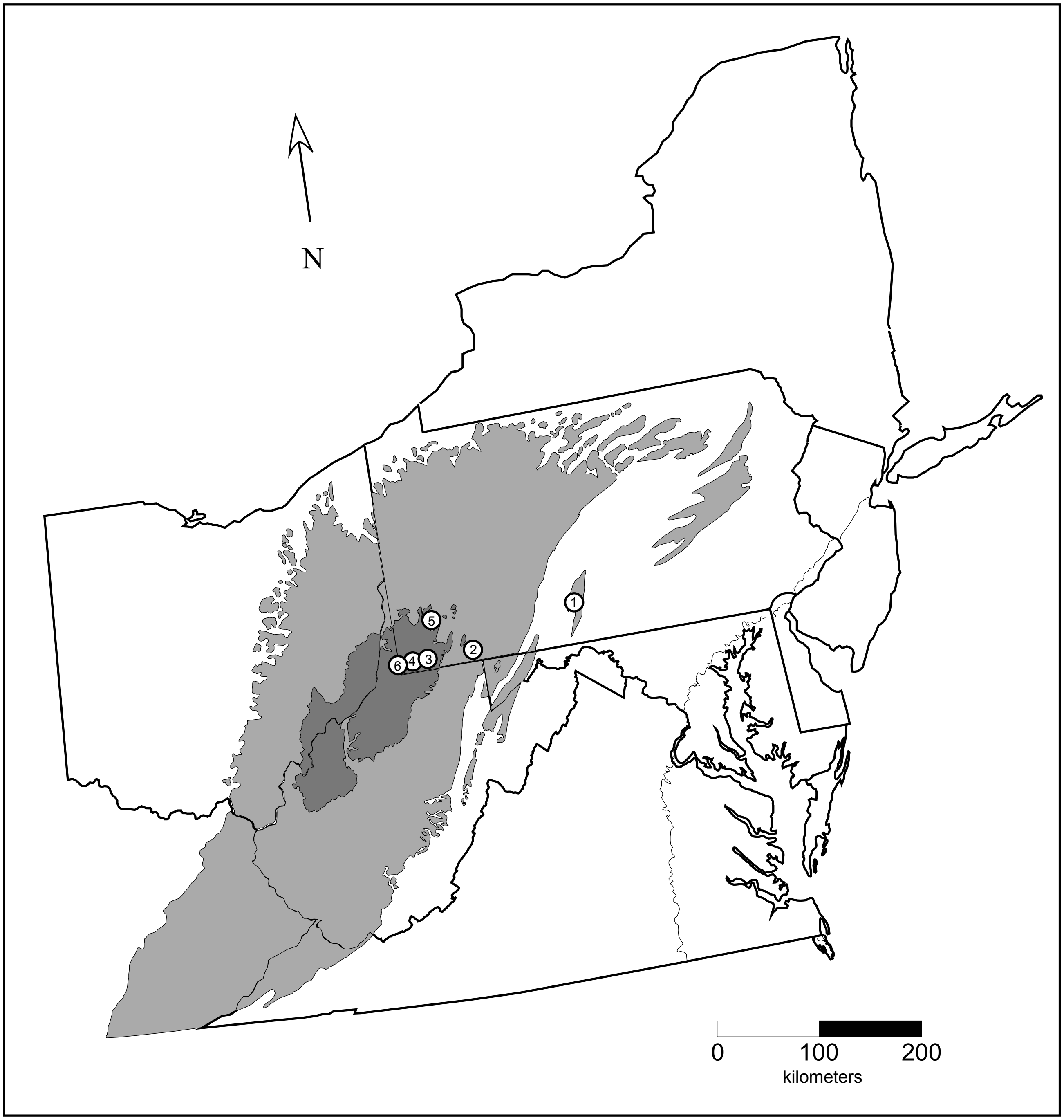

Figure 4.3. Locations of freshwater limestone samples from southwestern Pennsylvania. Pennsylvanian-age sedimentary deposits are shaded in light gray. Permian-age deposits are shaded in dark gray. Sample 1: Upper Freeport (?) limestone, Broad Top basin; Sample 2: Johnstown limestone, Forbes Forest; Sample 3: Upper Freeport limestone, Craynes Run; Sample 4: Little Pittsburgh limestone, Waynesburg; Sample 5: Benwood #1 & #2 limestones, Washington County; Sample 6: unnamed limestone in Washington Formation,Greene County.

climate change? Stable isotopic analysis of the lacustrine limestones in the Appalachian basin can be used to resolve whether or not the depositional environment was moist and humid (Valero-Garces et al., 1997) or increasingly arid (Cecil, 1990).

Lacustrine Limestones

Freshwater limestones have been observed in temperate (e.g. Eggleston and Dean, 1976; Carroll et al., 1983; Albarede and Michard, 1987; Drummond et al., 1993), subtropical (e.g. Monty and Hardie, 1976), and tropical environments (Hillaire-Marcel and Casanova, 1987), and their presence does not have unique environmental significance. They are typically deposited in lacustrine environments, and are common in seasonally wet ponds or shallow lakes that are inhabited by blue-green algae (Monty and Hardie, 1976; Eggleston and Dean, 1976). Several parameters may be responsible for the precipitation of calcite in these lakes, including the concentration of dissolved CO_2 and surface-water temperature, factors that are both physiochemical and biochemical in origin (e.g. Duston et al., 1986). Fine crystals of calcite are likely precipitated by the blue-green algal mat that may be aided by photosynthetic uptake of CO_2 and thermally induced density stratification of lake waters. The calcite is commonly re-crystallized by bacterial degradation of dead algal filaments (Monty and Hardie, 1976).

Although the depositional environment of the Pennsylvanian-Permian freshwater limestones is still a subject of some debate (e.g. Valero Garces et al., 1997; Cecil, 1990) derivation from a freshwater source generally is accepted. Within the Appalachian basin, the freshwater carbonates generally are closely associated with economic deposits of coal. This association is attributed to variations in local hydrological conditions that result from only a few tens of centimeters of topographic relief (Monty and Hardie, 1976). In areas of pervasive standing water, freshwater peat swamps thrive after aquatic plants have displaced the blue-green algae. By comparison, algal-lime mud deposits are produced in areas prone to temporary episodic flooding by freshwater. These depositional environments are not favorable for many freshwater plants. Freshwater peat and blue-green calcareous algal deposits are, therefore, antipathetic (Monty and Hardie, 1976). Either facies represents a depositional environment that is largely devoid of clastic input.

Stable Isotopic Approaches to Paleoclimatic Reconstructions

Carbonate lacustrine systems have the potential to yield critical paleoclimatic information by stable isotopic analyses. Lacustrine systems may preserve carbonate that formed in a hydraulically open or closed system. Open systems are characterized by continuous freshwater recharge and mixing that is balanced by losses to evaporation, transpiration, or outflow (Talbot, 1990). Hydraulically closed systems, by comparison, may persevere by episodic recharge, but remain isolated from the regional hydraulic system while evaporitic or transpiritic processes persist (Talbot, 1990). Freshwater algal carbonate mud is commonly deposited in seasonally wet environments (Monty and Hardie, 1976).

Both $\delta^{18}0$ and $\delta^{13}C$ of primary carbonate from lacustrine systems can be used to identify whether or not the hydraulic system behaved as an open or closed one. Open-system lakes show no covariance between the $\delta^{18}0$ and $\delta^{13}C$ of primary carbonate, because the carbon and oxygen pools are continuously replenished by groundwater, precipitation, or surface waters. As a result, water in an open-system lake has a very short residence time, and the carbonate composition may be used to calculate the composition of regional precipitation (Talbot, 1990). Closed-system lakes, in contrast, may display a pronounced covariance between the $\delta^{18}0$ and $\delta^{13}C$ of primary carbonate (Talbot, 1990). This happens because of evaporative loss of isotopically light water (Rayleigh fractionation), which enriches the concentration of ^{18}O in the lake. Carbonate precipitation, in contrast, preferentially removes the heavier ^{13}C isotope, so that a closed system evolves to more negative $\delta^{13}C$ values through time. However, the $\delta^{13}C$ values of the carbonate generally are more depleted during the wet season, and much more positive during the dry season, because of the dilution of the carbon pool by the influx of light organic carbon during times of increased biological activity (Gleason, 1972; Monty and Hardie, 1976). Citing studies of modern lacustrine systems, Talbot (1990) suggested that a correlation coefficient value $r > 0.7$ of $\delta^{13}C$ and $\delta^{18}0$ is typical of closed-system-lake environments. If covariance of the stable carbon and oxygen composition of lacustrine carbonate can be established, several additional paleoenvironmental inferences can be drawn, such as the surface area/volume of the lake (dependent on the slope of $\delta^{13}C/\delta^{18}0$ of the carbonate), and an estimate of the composition of precipitation (the most negative $\delta^{18}0$ value) (Talbot, 1990).

In this study, lacustrine limestone beds deposited at five different stratigraphic levels from the early-middle Pennsylvanian to the early Permian were studied to compare their stable isotopic compositions. Systematic changes in the $\delta^{18}O$ of the freshwater limestones may reflect changes in the precipitation resulting from the closing of the hypothetical Rheic Ocean during the initial stages of the Alleghanian orogeny (Faill, 1998). It is hypothesized that the development of Alleghanian orography fundamentally changed the source of precipitation to the Appalachian highlands from the Rheic Ocean to an epieric oceanic source from the continental interior of Laurentia.

Sampling Locations

Seven limestone samples from five stratigraphic levels in southwestern Pennsylvania were analyzed for their isotopic composition (Fig. 3). Pennsylvanian-Permian paleogeographic and paleoenvironmental reconstructions suggest that the drainages flowed from southeastern highlands toward an epieric sea to the west-northwest (Edmunds et al., 1979, 1999; Martin, 1998). Freshwater limestone deposition associated with ephemeral lakes began in the middle Pennsylvanian in southwestern Pennsylvania (Weedman, 1989) and continued through the Permian (Martin, 1998; Edmunds, 1999). The freshwater limestones are not formally recognized as members or formations, but several beds are recognized by the Pennsylvania and Ohio Geological Surveys as regionally correlative beds (Fig. 2). Their nomenclature will be used throughout this chapter, where appropriate, for the purpose of identifying stratigraphic position.

Many of the limestone samples listed below contain small calcium carbonate worm-tube casts of the polychaete worm genus Spirorbus. Cassle et al. (2003) point out that modern Spirorbids live in brackish water environments, and infer that their existence may be used as a criterion for distinguishing between brackish and freshwater conditions. Comparing the stable isotope and $^{87}Sr/^{86}Sr$ (Chapter 5) composition of the limestones will test this hypothesis.

Johnstown limestone (Allegheny Formation)

The middle Pennsylvanian Johnstown limestone lies beneath the Upper Kittanning coal (using nomenclature of Pennsylvania Geological Survey) of southwestern Pennsylvania (Weedman, 1989; Edmunds et al., 1999). It is the stratigraphically lowest freshwater limestone of any considerable extent in the northern Appalachian basin (Weedman, 1989). The hard, dark gray Johnstown limestone sample was collected from a core from Forbes Forest southeast of Uniontown, in Fayette County, Pennsylvania. The sample is a skeletal wackestone/boundstone, composed of very fine micritic mud, sub-millimeter bioclastic allochems of Spirorbis and ostracodes, and thin stromatolites. A few vertical fractures within the sample are filled with secondary calcite and interpreted to be desiccation cracks.

Upper Freeport limestone (Allegheny Formation)

The Upper Freeport limestone underlies the Upper Freeport coal seam, which marks the top of the Allegheny Formation (Edmunds et al., 1999). The Upper Freeport limestone is among the most studied of the freshwater limestones (Weedman, 1989, 1994; Valero-Garces, 1997). Two samples of Upper Freeport limestone were processed for stable isotopic analysis, although one sample has not been correlated by biostratigraphic means.

One sample is from a core that was drilled along Craynes Run, a small tributary located north-northeast of Waynesburg, Greene County, Pennsylvania. The sample is a dark brown, hard, and compact skeletal wackestone-packstone. Parts of the sample appear to be clast-supported, but most of the rock is composed of ostracod and Spirorbus allochems. The sample also contains very fine disseminated pyrite.

Another sample is from a Pennsylvania Geological Survey core hole, in the Saxton, Pennsylvania, 7.5-minute quadrangle, at a depth of 544.5 feet (1519.5 feet above sea level). The sample is considered approximately equivalent to the Upper Freeport limestone, although correlation is complicated by a lack of correlative beds that can be related to the well-established Pennsylvanian-age stratigraphy of southwestern Pennsylvania. The sample site is in the Broad Top basin in central Pennsylvania, which is east of the mappable limit of the Upper Freeport limestone. The sample is a compact, grayish-brown lime mudstone, with suspended silt-sized detrital quartz. Some small cryptalgal structures (oncolites?) within the sample contain coarser quartz silt. No bioclastic allochems could be identified in the sample.

Lower Pittsburgh limestone (Casselman Formation)

The Lower Pittsburgh limestone is just below the Little Pittsburgh coal seam near the top of the Casselman Formation of the Conemaugh Group in the western Pennsylvania coal field (Edmunds et al., 1999). The sample was collected from a core in Waynesburg, Greene County, Pennsylvania. The 4-inch thick, hard, compact, dark, grayish-brown limestone is about 30 feet below the Pittsburgh coal seam. It may be equivalent to the Lower Pittsburgh limestone. The sample is most easily classified as a skeletal wackestone; but in detail, it is a mixture of oncolites and cryptalgal structures, bioclasts of ostracodes and Spirorbis, and silt-sized calcitic spar.

Benwood Limestone (Monongahela Group)

The Benwood limestone is within the Pittsburgh Formation of the Monongahela Group and can be traced from southwestern Pennsylvania, into southeastern Ohio and northwestern West Virginia. The depositional environment of the Benwood has been studied in detail by Petzold (1988, 1989), who classified it into three main facies: 1) a deep-water lacustrine to terrestrial sequence, 2) a shallow lake to shoreline carbonate-siliciclastic sequence, and 3) a lacustrine to arid mudflat sequence. Glascock and Gierlowski-Kordesch (2002) interpret the Benwood limestone in Ohio to be an open lake system that mixed intermittently with marine waters.

Two samples of the Benwood limestone were collected for analysis from Pennsylvania Department of Transportation core holes. Both samples are from the Hackett 7.5' quadrangle in Washington County. At one location (Benwood #1), the Benwood limestone is light gray, hard, compact, and approximately 1.2 meters (4 feet) thick. In the other core (Benwood #2), the Benwood limestone is light gray, hard, and compact, but only 0.85 meters (2.8 feet) thick. The Benwood #1 sample is a micritic boundstone composed of oncolites, with rare rounded polycrystalline quartz silt and disseminated fine sub-millimeter pyrite. The Benwood #2 sample is a lime mudstone with sparry pore fillings that are interpreted to be diagenetic in origin. Rare well-rounded, silt-sized, polycrystalline quartz grains suspended in the micritic mud are likely eolian in origin. The sample also includes some very fine (sub-millimeter) disseminated pyrite.

Unnamed Limestone in the Washington Formation (Dunkard Group)

The stratigraphically highest freshwater limestone analyzed is from a core from Greene County, 235 meters (770 feet) above the Pittsburgh coal, which marks the base of the Monongahela Group. This places the limestone in the middle of the Washington Formation of the Dunkard Group. The limestone in the core is tan to gray, hard, compact, and about 0.15 meters (0.5 feet) thick. It is a skeletal wackestone with sub-millimeter allochems of ostracodes and Spirorbis and very fine disseminated pyrite.

Methods

Bulk samples of the Upper Freeport (?) from Broad Top basin, Lower Pittsburgh, Benwood #2, and unnamed Washington limestones were prepared for the purpose of establishing the range of $\delta^{18}O$ and $\delta^{13}C$. The samples were prepared by sampling carbonate at several places within the core with a 1 mm drill bit, and thoroughly mixing the powdered limestone.

Higher resolution sampling was achieved by drilling at five 1-cm intervals normal to bedding for $\delta^{18}O$ and $\delta^{13}C$ analysis. Stable isotopic analysis of the lacustrine limestones was performed at the Environmental Research and Training Laboratory (ERTL) at the University of Kentucky under the direction of Dr. Harold Rowe. Approximately 40 to 60 micrograms of sample was reacted with anhydrous phosphoric acid at 70^{o}C and allowed to react for ~2 hours prior to analysis. The evolved CO_2 was mobilized with a He carrier gas and dehydrated by passing through a Natheon tube (which diffuses out water, but not CO_2 from the gas stream). The purified gas was analyzed for mass 44, 45, 46 a.m.u. with a Finnegan Delta Plus sector mass spectrometer. Precision and accuracy were monitored by sequential analyses of the NBS-19 carbonate standard.

Results

Bulk samples of the freshwater limestones were analyzed following the same acidification method outlined above. Stable oxygen isotopic analyses of the bulk composite sample of the Upper Freeport (?) from Broad Top basin had an unacceptably high (>0.2 $^{o}/_{oo}$) standard deviation (Table 1), and may reflect the presence of detrital clay that might have reacted with the phosphoric acid. Because of this complication, additional high-resolution analyses were not attempted on this sample. A bulk composite sample of the Lower Pittsburgh sample was analyzed for $\delta^{18}0_{PDB}$ and $\delta^{13}C_{PDB}$ (PDB: Pee Dee Belemnite standard). The results of the bulk sample (Table 1) are generally similar to those obtained from higher resolution sampling from the same core (Table 2). Micritic mud matrix and sparry cavity fillings of the Benwood #2 sample were sampled individually to see if the isotopic composition varied. The sparry cement of the Benwood #2 sample has a much more negative $\delta^{18}0_{PDB}$ than the micrite, although the $\delta^{13}C_{PDB}$ is similar in both textures (Table 1). As a result, the Benwood #2 sample was not analyzed for covariance and for determination of the isotopic composition of paleo-meteoric water in order to avoid mixing multiple generations of calcite. A bulk sample of the Washington unit was also analyzed for $\delta^{18}0_{PDB}$ and $\delta^{13}C_{PDB}$ (Table 1).

The freshwater limestone samples have $\delta^{18}0_{PDB}$ values ranging from -6.4 to 0.05$^{o}/_{oo}$ and $\delta^{13}C_{PDB}$ values from -0.4 to -9.5 $^{o}/_{oo}$(Table 2). The data were plotted on a chart with $\delta^{13}C_{PDB}$ on the ordinate axis and $\delta^{18}O_{PDB}$ on the abscissa. Linear regression on the $\delta^{13}C_{PDB}$ and $\delta^{18}O_{PDB}$ plots was performed to determine if the data display any covariant trends. Equations of best-fit linear regression lines and r^2 correlation coefficients are displayed in Table 2.

Sample ID	d^{13}C$_{PDB}$	d^{13}C$_{stdev}$	d^{18}O$_{PDB}$	d^{18}O$_{SMOW}$	d^{18}O$_{stdev}$
unnamed Washington	-6.53	0.10	0.22	31.14	0.16
Benwood #2 spar	-5.39	0.08	-6.24	24.49	0.16
Benwood #2 mic	-4.98	0.12	4.76	35.83	0.11
Lower Pittsburgh	-8.52	0.08	-0.16	30.75	0.14
Upper Freeport (?)	-1.26	0.09	-7.57	23.12	0.24

Table 4.1. Stable isotopic data of bulk samples for selected lacustrine limestones from the Appalachian basin in southwestern Pennsylvania. The data are arranged with the youngest samples at the top, getting progressively older toward the bottom. (PDB: Pee Dee Belemnite; SMOW: Standard Mean Ocean Water)

Sample ID	$\delta^{13}C_{PDB}$	$\delta^{13}C_{stdev}$	$\delta^{18}O_{PDB}$	$\delta^{18}O_{SMOW}$	$\delta^{18}O_{stdev}$	Lake Water ($\delta^{18}O_{SMOW}$) 25°C	30°C

unnamed Washington limestone

W-1	-4.68	0.06	-3.52	27.29	0.08	-1.09	-0.07
W-2	-6.76	0.12	-4.71	26.06	0.11	-2.32	-1.30
W-3	-6.72	0.09	-5.13	25.63	0.13	*-2.75*	*-1.73*
W-4	-6.32	0.11	-4.07	26.72	0.05	-1.66	-0.64
W-5	-6.21	0.077	-3.88	26.92	0.06	-1.47	-0.44

$\delta^{13}C/\delta^{18}O$ covariance: $r^2=0.67$ r=0.82 equation: $\delta^{18}O = (0.6247*\delta^{13}C)-0.4284$

Benwood #1 limestone

B-1-1	-0.77	0.06	0.50	31.43	0.07	*3.05*	*4.07*
B-1-2	-0.43	0.05	0.50	31.43	0.10	*3.05*	*4.07*
B-1-3	-0.64	0.07	0.05	30.97	0.09	*2.58*	*3.61*
B-1-4	-0.56	0.10	0.39	31.33	0.08	*2.94*	*3.97*
B-1-5	-0.67	0.10	0.35	31.28	0.06	*2.90*	*3.92*

$\delta^{13}C/\delta^{18}O$ covariance: $r^2=0.02$ r=0 equation: $\delta^{18}O = (0.2118*\delta^{13}C)+0.4865$

Pittsburgh limestone

P-1	-8.51	0.09	-5.59	25.16	0.09	-3.22	-2.20
P-2	-9.44	0.11	-6.06	24.67	0.06	*-3.72*	*-2.69*
P-3	-8.48	0.10	-5.13	25.63	0.06	-2.75	-1.73
P-4	-6.70	0.07	-3.88	26.92	0.05	-1.46	-0.44
P-5	-7.79	0.08	-3.70	27.10	0.11	-1.28	-0.26

$\delta^{13}C/\delta^{18}O$ covariance: $r^2=0.78$ r=0.88 equation: $\delta^{18}O = (0.906*\delta^{13}C)+2.5412$

Upper Freeport limestone

UF-1	-2.13	0.10	-4.87	25.90	0.10	-2.49	-1.46
UF-2	-2.04	0.06	-4.79	25.98	0.08	-2.40	-1.38
UF-3	-1.87	0.09	-5.06	25.70	0.05	*-2.68*	*-1.66*
UF-4	-1.26	0.06	-4.39	26.40	0.10	-1.98	-0.96
UF-5	-0.71	0.06	-3.72	27.09	0.06	-1.29	-0.27

$\delta^{13}C/\delta^{18}O$ covariance: $r^2=0.88$ r=0.94 equation: $\delta^{18}O = (0.8319*\delta^{13}C)-3.2334$

Johnstown limestone

J-1	-4.17		-6.12	24.61		*-3.77*	*-2.75*
J-2	-3.81		-5.94	24.80		*-3.59*	*-2.56*
J-3	-3.84		-6.21	24.52		*-3.87*	*-2.84*
J-4	-4.54		-5.94	24.80		*-3.59*	*-2.56*
J-5	-4.66		-6.37	24.35		*-4.03*	*-3.01*

$\delta^{13}C/\delta^{18}O$ covariance: $r^2=0.10$ r=0.32 equation: $\delta^{18}O = (0.1504*\delta^{13}C)-5.4839$

Table 4.2. Stable isotopic data from the Pennsylvanian-Permian freshwater limestones. Covariant trends were determined by linear regression. Lake water values (best estimates in italics) were calculated using the empirical fractionation equations of O'Neil et al. (1969). Stratigraphically youngest samples are listed at the top, getting progressively older toward the bottom.

Discussion

The average $\delta^{18}O$ of the limestones does not show any consistent depletion through time that suggested the progressive development of an Alleghanian rain shadow. The $\delta^{18}O$ of the lake waters may have been influenced by isotopic exchange during mineral dissolution in the drainage network that would have enriched the $\delta^{18}O$ of the bicarbonate ion, or perhaps the composition of the source of water vapor was unaffected by the uplift of the Appalachians.

The $\delta^{18}0$ and $\delta^{13}C$ of three of the five samples (Upper Freeport, Pittsburgh, and Washington limestones) show a statistically significant covariant trend of $r>0.7$, indicating that they were precipitated in an isotopically closed system (Table 2). The Johnstown and Benwood #1 limestones do not exhibit any covariant behavior, indicating that they were deposited in an isotopically open system. The small variation (Table 2) in $\delta^{18}0$ and $\delta^{13}C$ between the Johnstown and Benwood #1 samples indicates that the waters in the lacustrine system were of fixed composition, had a short residence time, and therefore, can be used reliably to approximate the composition of regionally averaged precipitation (Talbot, 1990). The isotopic composition of precipitation in the catchment areas of the closed-system lakes can be approximated from the isotopically lightest carbonate analyses of these systems (Talbot, 1990). This is a reasonable assumption because evaporative loss tends to drive the lake waters toward an isotopically heavier composition.

In order to have a convenient reference frame for interpretation of $\delta^{18}0$ isotopic data from the carbonates, the values were converted from the PDB standard to the SMOW (Standard Mean Ocean Water) standard. This was done by the conversion:

$$\delta^{18}0_{SMOW} = 1.03091 * \delta^{18}0_{PDB} + 30.91.$$

Because the oceans are the largest reservoir of water on the planet, and because most precipitation is water derived from evaporation of ocean water, it is reasonable and more intuitive to use sea-water isotopic composition ($\delta^{18}O_{SMOW}$=0) as a starting point.

The $\delta^{18}0$ of the precipitation in equilibrium with the Pennsylvanian-Permian freshwater limestones was done by using the stable isotope fractionation factors of O'Neil et al. (1969).

$$1000*\ln \alpha_{calcite\text{-}water} = 2.78 * (1x10^6/T^2)-2.89,$$

where $\alpha_{calcite\text{-}water}$=fractionation factor between calcite-water

T=temperature (Kelvin).

To calculate the composition of meteoric waters, it is necessary to know the fractionation factor of $^{18}O/^{16}O$ between the two phases in equilibrium. Fractionation is a temperature dependent process depending on molecular translational velocity, vibrational energy of bonds, and associated free energy of formation. Therefore, a temperature must be assumed for the lake waters. Because the Appalachian basin was located in the equatorial latitudes during the Pennsylvanian and Permian (Cecil, 1990), two temperatures of 25^{o}C and 30^{o}C were used to bracket the estimated range of temperatures of the lake waters to calculate the composition of the $\delta^{18}0_{water}$.

Following that fractionation is defined by:

$$\delta^{18}0_{calcite}\text{-}\delta^{18}0_{water}=\Delta^{18}0_{calcite\text{-}water},$$

and that $\Delta^{18}0_{calcite\text{-}water}$ is a close approximation of (1000 * ln $\alpha_{calcite\text{-}water}$), the isotopic composition of waters in equilibrium with the calcite can be calculated (Table 2).

Note that there is no accounting for a "vital effect" of enhanced isotopic fractionation by blue-green algae. A single temperature estimate for the lakes is probably reasonable, considering that the origin of the freshwater carbonate is likely linked to precipitation out of the surface epilimnion or from photosynthetic blue-green algae (Talbot, 1990). Because the Appalachian basin occupied a position near the equator, neither seasonal temperature differences nor compositional changes from vertical mixing within the lake are likely significant.

The $\delta^{18}O_{SMOW}$ values for precipitation (Table 2) are equal to, or enriched in ^{18}O compared to the values determined for late Paleozoic seawater ($\delta^{18}O_{SMOW}$= -3 to -1; Mii et al., 1999). Evaporative processes always result in a depletion of ^{18}O in the vapor phase, and there is no mechanism that will result in meteoric waters that are isotopically heavier than their source. The $\delta^{18}O$ of the lake waters must have been enriched by evaporative loss of ^{16}O. This interpretation is supported by the covariant behavior of $\delta^{13}C$ and $\delta^{18}O$, so that these limestones are interpreted to have been deposited in an evaporative, closed lake system that was seasonally refreshed with a combination of groundwater and/or meteoric waters. The lack of covariance in the $\delta^{13}C$ and $\delta^{18}O$ in the Johnstown and Benwood limestones might reflect continuous recharge of isotopically enriched groundwater that kept the system well mixed.

The $\delta^{13}C_{PDB}$ of the five limestone samples are considerably lighter (-0.43 to -9.44) than the $\delta^{13}C$ of early Paleozoic marine carbonates ($\delta^{13}C_{PDB}$> 0; Keith and Weber, 1964) that might have been exposed in the sedimentary source regions. This suggests that oxidized organic matter was a likely contributor to the dissolved inorganic carbon pool in these lakes. The source of organic matter in the Appalachian basin during the Pennsylvanian-Permian is assumed to be peat.

The slope of the covariant $\delta^{13}C$ and $\delta^{18}O$ trend in the Upper Freeport, Pittsburgh, and Washington limestones may also be used to infer the paleoenvironmental setting of these lacustrine systems. Talbot (1990) hypothesized that evaporative lakes with high surface area:depth ratios will have shallow $\delta^{13}C$ and $\delta^{18}O$ covariant slopes, whereas lakes with low surface area:depth ratios will have steep $\delta^{13}C$ and $\delta^{18}O$ covariant slopes. The Upper Freeport, Pittsburgh, and Washington limestones have covariant slopes (Table 2-equations) that are <1, suggesting that they are relatively shallow.

In a previous study of the Middle Pennsylvanian Upper Freeport freshwater limestone, Valero-Garces et al. (1997) interpreted the relatively depleted $\delta^{18}O$ values as evidence that the freshwater limestone was deposited in a non-arid climate. Following the criteria of Talbot (1990) to discriminate between closed- and open-system lacustrine environments, Valero-Garces et al. (1997) also showed that there is no systematic covariation of $\delta^{13}C$ and $\delta^{18}O$ values within the Upper Freeport limestone, which is evidence that the environment remained well mixed and open to exchange with meteoric water. Their estimate of the isotopic composition of the $\delta^{18}O$ of waters in equilibrium with the limestone ranged from -2.4 to -1.4 $^{o}/_{oo}$. The Valero-Garces et al. (1997) dataset for the Upper Freeport limestone is much more extensive than the one from this study. Although they did not observe a covariant trend with the $\delta^{13}C$ and $\delta^{18}O$ data, the disparity between their dataset and ours (Table 2) is likely a reflection of the ephemeral and discontinuous nature of these lacustrine systems. An isopach map of the Upper Freeport limestone in western Pennsylvania shows that it was not deposited as one continuous lake, but instead is composed of several systems that are isolated by fluvial levees and channels (Weedman, 1989). It is likely that the Upper Freeport limestone that we sampled was from more ephemeral lake system than the Valero-Garces et al. (1997) samples of Upper Freeport limestone. The overlap of $\delta^{13}C$ and $\delta^{18}O$ values in both datasets is additional assurance that the isotopic values are reproducible within a certain range.

The $\delta^{13}C$ and $\delta^{18}O$ values from the Pennsylvanian and Permian freshwater limestones support models of deposition in a semi-arid environment (e.g. Cecil, 1990). The appearance of freshwater limestones in the Allegheny Formation of southwestern Pennsylvania accompanies a change in the Pennsylvanian-age floral populations from varieties suited to humid environments to those suited to more arid, well-drained soils (Pfefferkorn and Thomson, 1982; Wagner and Lyons, 1997; Dimichele et al., 2001). The progressive increase in the number of paleosols throughout the late Pennsylvanian (Cecil, 1990) and into the Permian (Martin, 1998) is interpreted to represent deposition in a tropical climate that alternated between heavy rainfall and drought. Atmospheric circulation models, eolian cross-bed data, pedogenic mineral $\delta^{18}O$ compositions, and detrital-zircon populations from the western United States also suggest a change from zonal circulation in the middle Pennsylvanian to strongly monsoonal circulation in the Early Permian (Parrish, 1993; Soreghan et al., 2002; Tabor and Montanez, 2002). These conditions may have a stronger correlation to the growth of Pennsylvanian-Permian continental glaciers that have been documented throughout many of the Gondwanan continents (summarized in Caputo and Crowell, 1985) than the uplift of the

Appalachians. The stable isotopic data from this study support a seasonally arid paleoclimate during the middle Pennsylvanian-Permian.

Conclusions

The preliminary results from study of Pennsylvanian–Permian lacustrine carbonates support deposition in an arid environment (Cecil, 1990). The relatively heavy $\delta^{18}O_{SMOW}$ of lake waters in equilibrium with the Upper Freeport, Pittsburgh, Benwood, and Washington limestones is evidence that the lake waters were highly fractionated by Rayleigh distillation. The source of precipitation cannot be uniquely determined, however, it probably came from a western-southwestern epeiric sea within Laurentia (such as the Midland, Delaware, Paradox, or Eastern Interior Seas) because they were the closest marine water bodies to the Appalachian basin. Another possibility for the source of precipitation was the hypothetical Rheic Ocean along the eastern margin of Laurentia, unless the juxtaposition of Gondwana during the late Paleozoic had destroyed it by Pennsylvanian time.

It might be assumed that the Alleghanian orogen might have developed a topographic barrier that would have prevented precipitation from the Rheic Ocean to reach the Appalachian basin. Most models of the late Paleozoic paleoclimate of Pangea (Otto-Bleisner, 1993; Kutzbach, 1994; Golanka et al., 1994) rely on the assumption that the Alleghanian orogen had generated topography rivaling that of the modern Himalayas. However, is likely that the Alleghanian orogeny had not developed a high topographic expression through the early Permian. K/Ar ages of detrital white mica from the Pennsylvanian-Permian are all of Devonian age (Aronson and Lewis, 1994; Becker and

Aronson, unpubl. data), suggesting that the crust was not exhumed to significant depth. It is hypothesized that thermochronologic cooling ages from orogenic belts should evolve to ages that are synorogenic and of constant value as the orogeny proceeds (Willet and Brandon, 2002). The lack of Alleghanian synorogenic ages of the detrital white mica suggests that the topography was developed by significant lateral translation of crust. Similar patterns have been noted in Taiwan (Willet et al., 2003). The age of Alleghanian synorogenic clastic wedge deposits ranges from ~330 Ma to ~285 Ma (Ross and Ross, 1987; Edmunds et al., 1999). Over the 45 m.y. of orogenic activity, rocks were not exhumed from temperatures of 350^{o}C (depth of ~12 km), which indicates very low erosion rates and thus the very low rates of orographic development. Contrary to some models of Appalachian paleotopography (Slingerland and Furlong, 1989), it is hypothesized that the Appalachians of the Pennsylvanian and early Permian were relatively low, unlike the modern-day Himalayas.

Accomodation space to allow for the formation of carbonate lakes has also been hypothesized to result from the thrust-sheet loading along the eastern margin of Laurentia (Weedman, 1989). The propagation of thrusts into the Laurentian continental interior likely post-dated the deposition of the freshwater limestones in the middle Pennsylvanian (e.g. Miller and Kent, 1988; Stamatokos et al., 1996). The appearance of freshwater limestones in the stratigraphic record does not reflect a specific tectonic environment, but probably reflects a combination of regional climatic change and reduction of clastic input into the Appalachian basin from the middle Pennsylvanian through the early Permian. The relatively shallow depth of the lakes, and their tendency to be ephemeral depositional environments, suggests that the lakes are not attributed to foreland flexure, but rather a random consequence of deposition within peat swamps.

CHAPTER 5. EVOLUTION OF THE ALLEGHANIAN OROGEN DETERMINED FROM RADIOGENIC STRONTIUM IN LATE PALEOZOIC LACUSTRINE LIMESTONES FROM THE NORTHERN APPALACHIAN BASIN

Introduction

The Appalachian basin is a composite Paleozoic foreland basin formed in response to at least three major Paleozoic orogenic events. The last of these events, the late Paleozoic Alleghanian orogeny, was a collision between the Gondwanan and Laurentian continents that produced the Pangean supercontinent. This event (or series of events) is represented in the Appalachian basin by the progradation of a deltaic depositional system onto an epieric marine carbonate shelf in the late Mississippian and early Pennsylvanian (Thomas, 1974; 1977).

Mechanical and kinematic models of collisional orogens are designed to follow a critical taper geometry, where the tectonic imbrication of continental crust begins nearest to the edge of continental plate and advances toward the craton in a break-forward sequence (Davis et al., 1983). This results in the progressive and rapid exhumation of rocks from considerable depth near the collisional plate boundary.

Studies of shear zones within the Alleghanian collisional orogen, however, suggest that most of the early deformation was transpressional (Gates et al., 1986, 1988; Krol et al., 1999; Valentino and Gates, 2001; Hatcher, 2001). Propagation of craton-directed thrusts into the foreland did not occur until the latest Pennsylvanian in the southern Appalachians, and the middle-late Permian in the central Appalachians (Miller and Kent, 1988; Stamatakos et al., 1996).

A critical part of the tectonic history of the Alleghanian orogeny is preserved within the late Mississippian-early Permian sedimentary record of the Appalachian basin.

This record is useful in determining the crustal composition of the orogen during the initial stages of the continental collision. The exhumational history of the Alleghanian orogeny has been inferred from sedimentary petrography (Dickinson et al., 1983; Davis and Ehrlich, 1974; O'Connor, 1989) and applications of thermochronological and geochronological proxies for the Pennsylvanian and Permian strata preserved within the Appalachian basin (e.g., Aronson and Lewis, 1994; Gray and Zeitler, 1997; Thomas et al., 2004a; Chapter 2). The discovery of Alleghanian-age dextral shear zones opens the possibility that the early-middle Pennsylvanian clastic deposits are a record of transpressional uplift of Laurentian crust (Chapter 2.), and not an orthogonal continental collision.

Petrographic analyses of lower-middle Pennsylvanian sandstones from the central Appalachian basin are reported to reveal a sequence of unroofing, beginning with a recycled sedimentary and low-grade metamorphic source to exposure of a granitic or crystalline basement sedimentary source (Davis and Ehrlich, 1974; O'Connor, 1989). Following these observations, it can be inferred that the source of sediment was not intensely altered by weathering processes, and that the middle Pennsylvanian sediment was in a first cycle of erosion and deposition. In contrast, the petrography of the upper Pennsylvanian-lower Permian Dunkard Group (preserved in the northern Appalachian basin) is interpreted to reflect a source from the southeast that was dominated by low-grade metamorphic and recycled sedimentary rocks (Martin, 1998). Martin's (1998) data would suggest that the low-grade parts of the orogen continued to be the dominant source of sediment to the Appalachian foreland basin. The sedimentary record in the central Appalachian basin (Pocahontas, New River, and Kanahwa Formations) is likely part of the same pervasive drainage system that provided sediment to the Dunkard Group of the northern Appalachian basin (Thomas, 1977; Edmunds et al., 1979; Martin, 1998). An analysis of the sedimentary record in this part of the Appalachian basin might yield additional information about the progressive deformation of the Laurentian margin during the late Paleozoic.

K-Ar dating of bulk detrital white mica separates from Pennsylvanian-age to Permian-age sandstones within the Appalachian basin reveals that the source of sediment was exhumed from depths below the closure temperature of white mica (350^o +/- 50^oC, McDougall and Harrison, 1999) at ~360 Ma in the early Pennsylvanian and ~400 Ma in the late Pennsylvanian/early Permian (Aronson and Lewis, 1994; Becker and Aronson, unpub. data). Although it can be argued that the bulk white mica K-Ar ages from the Pennsylvanian and Permian deposits may not be sensitive to the successive addition of younger constituents, $^{40}Ar/^{39}Ar$ analyses of single white mica crystals from the upper Pennsylvanian-age Llewellyn Formation in eastern Pennsylvania also reveal no ages younger than Devonian (P. Zeitler and B.Idleman, pers. comm.). This implies that Alleghanian synorogenic sediment supplied to the Appalachian basin was from a source not exhumed from a depth greater than 10-12 kilometers (assuming a geothermal gradient of 30^oC/km).

U-Pb ages of detrital zircons from the early to middle Pennsylvanian sandstones suggest that most of the detritus within the Appalachian basin was recycled from Mesoproterozoic basement and late Proterozoic and Paleozoic strata of the Laurentian margin (Gray and Zeitler, 1997; Thomas et al., 2004a; Eriksson et al., 2004; Chapter 2). Archean and late Paleoproterozoic (1600-1900 Ma) age detrital zircons are cited as evidence of recycling of the Laurentian syn-rift and passive-margin sandstones (Thomas et al., 2004a). Detrital zircon ages from early-middle Permian-age sandstones of the Dunkard Group do not contain any Archean detrital zircon ages. Instead, the detrital-zircon ages from the Dunkard Group support a source of sediment with a much more restricted age population, possibly the igneous and metamorphic internides or middle Paleozoic sandstones from the Appalachian basin (e.g., Bream, 2002; Eriksson et al., 2004). Detrital-white-mica ages and detrital-zircon ages from Pennsylvanian-age sandstones indicate that the late Paleozoic orogen did not incorporate any significant synorogenic juvenile crust. Detrital- zircon ages from Permian sandstones of the Dunkard Group suggest either a progressive unroofing of the orogenic highland into crystalline rocks of metamorphic and igneous origin and/or recycling of Paleozoic sedimentary deposits.

Lacustrine carbonates within the Pennsylvanian-age stratigraphy offer yet another opportunity to evaluate the sequence of tectonic assemblage of the Alleghanian orogeny. These carbonates are discontinuous thin lenses within the Pennsylvanian-age Allegheny, Conemaugh, and Monongahela Formations and the early Permian-age Washington and Greene Formations. The carbonates provide a rare opportunity to sample the geochemical characteristics of the waters that drained the late Paleozoic orogen, including the isotopic composition of the precipitation and a bulk sampling of the dissolved $^{87}Sr/^{86}Sr$ contributed from weathering of silicate and carbonate rocks that comprise the orogen. Therefore, the freshwater carbonates may afford a unique opportunity to evaluate the sequence of tectonic assemblage of the Pangean supercontinent during the middle Pennsylvanian through early Permian stages of the late Paleozoic collision.

Strontium Isotopes in Lacustrine Limestones

The $^{87}Sr/^{86}Sr$ ratio of river waters from drainages crossing a mountain belt would be a composite of the diverse radiogenic Sr sources that are exposed during an orogenic event. However, in order to develop a kinematic history using Sr, the different elements that compose the origin must be distinctive. Riverine $^{87}Sr/^{86}Sr$ ratios have been suggested to be buffered by the weathering of carbonates which have a high concentration of strontium (Brass, 1976). On the basis of mass balance experiments, Brass (1976) concluded that roughly 75% of riverine strontium entering the world's oceans is derived from weathering of carbonates. Long-term variations in the $^{87}Sr/^{86}Sr$ ratio of radiogenic strontium in freshwater limestones are likely to reflect sources from weathered silicate minerals because Paleozoic marine limestones have a restricted range of $^{87}Sr/^{86}Sr$ ratios (e.g., Burke et al., 1982; Veizer et al., 1999), whereas silicate minerals have a very large range of $^{87}Sr/^{86}Sr$ ratios. It might be expected, therefore, that the tectonic evolution of an orogenic belt might be reflected in the $^{87}Sr/^{86}Sr$ of freshwater limestones, as various crystalline components of parts of the orogenic belt are exhumed and exposed to erosion.

Sediments are typically derived from weathering profiles (Nesbitt et al., 1996), but not all silicate rock-forming minerals weather at the same rate (e.g., White et al., 2001), and not all of them have the same $^{87}Sr/^{86}Sr$ ratios or concentration of Sr, because of elemental partitioning of different mineral phases during crystallization from a silicate melt. Some minerals (such as olivine, plagioclase, hornblende, and clinopyroxene) incorporate very little rubidium into their lattice during crystallization, so that the initial

$^{87}Sr/^{86}Sr$ ratios do not change much through time. In contrast, biotite is one of the largest reservoirs of radiogenic strontium of the common rock-forming minerals, because of a relatively high concentration of Rb and low concentration of Sr, which follows the partitioning of K over Ca during crystallization of biotite in the parent melt. With time, the $^{87}Sr/^{86}Sr$ ratio increases rapidly. Blum and Erel (1997) demonstrated that the $^{87}Sr/^{86}Sr$ ratio of soils decreases as a function of increasing exposure, and that it was because of the weathering of biotite. Biotite rapidly alters to hydrobiotite or vermiculite, releasing elevated $^{87}Sr/^{86}Sr$ (Blum and Erel, 1997) that is dissolved into groundwater. In more arid to temperate environments, biotite first weathers to hydrobiotite (mixed vermiculite-biotite) before transforming completely to vermiculite. In hot, humid environments, the biotite-vermiculite reaction proceeds much more rapidly, and may further transform to kaolinite (Blum and Erel, 1997). Therefore, during rapid exhumation of an orogen that involves progressive exposure of highly radiogenic rocks, it might be expected that the strontium in waters draining the region would gradually become more radiogenic through time, until tectonism subsided and the landscape developed greater antiquity.

Global seawater $^{87}Sr/^{86}Sr$ ratios vary non-systematically through time. Relatively high concentrations of radiogenic strontium were dissolved in the marine waters through the late Carboniferous and Early Permian (Denison et al., 1994; Veizer et al., 1999) suggesting enhanced weathering of old continental crust that may be linked to climatically moderated chemical weathering and tectonism (Richter et al., 1992) or to continental glaciation (Caputo and Crowell, 1985). Decreases in the concentration of radiogenic strontium in seawater are caused by enhanced hydrothermal alteration of oceanic basalt (Spooner, 1976; Albarede et al., 1981). The question as to whether increases in radiogenic strontium are caused by the massive influx of interglacial sediment, or by enhanced chemical weathering of old continental crust remains open to debate. Present-day sedimentary budgets indicate that although the tropics occupy only 25% of the Earth's land area, they contribute 65% of the dissolved silica load and 38% of the ionic load (Stallard, 1992). During the late Paleozoic, eastern Laurentia occupied a low-latitude position (Otto-Bleisner, 1993), suggesting that weathering of the tropical Appalachians must have been a very important contributor of radiogenic strontium to the world's oceans.

The $^{87}Sr/^{86}Sr$ ratio of freshwater limestones in the Appalachian basin should help to resolve 1) whether or not these limestones are indeed of freshwater origin or open to exchange with marine water (Cassel et al., 2003); 2) the contribution of radiogenic ^{87}Sr from the Alleghanian orogen to the global seawater for the late Paleozoic (Denison et al., 1994); and 3) the exhumational history of disparate components of the Alleghanian orogeny.

Sampling Locations

The sampling locations for the freshwater limestones that were analyzed in this study are identical to those for stable isotopic analysis (Chapter 4), with the exception of one additional sample of Upper Freeport limestone (Fig. 1). The additional sample of the Upper Freeport limestone was collected from the New Portage Railroad Tunnel east of Gallitzen, Pennsylvania, very near the Cambria-Blair County line. Similar to the core sample of the Upper Freeport from Craynes Run, this outcrop sample is dark gray, hard, compact, skeletal wackestone, containing small casts of Spirorbus.

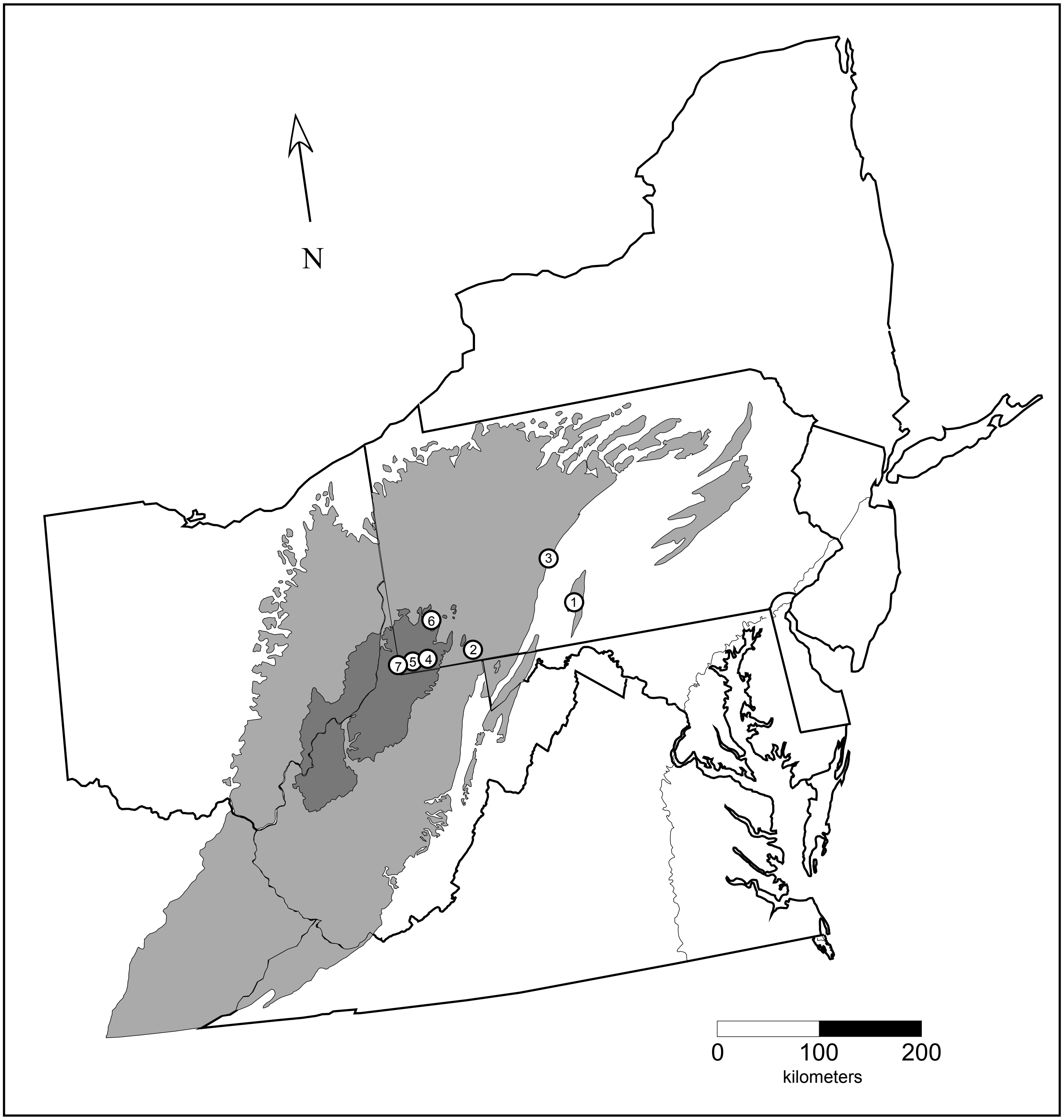

Figure 5.1. Locations of freshwater limestone samples from southwestern Pennsylvania. Pennsylvanian-age sedimentary deposits are shaded in light gray. Permian-age deposits are shaded in dark gray. Sample 1: Upper Freeport (?) limestone, Broad Top basin; Sample 2: Johnstown limestone, Forbes Forest; Sample 3: Upper Freeport limestone, Gallitzen; Sample 4: Upper Freeport limestone, Craynes Run; Sample 5: Little Pittsburgh limestone, Waynesburg; Sample 6: Benwood #1 & #2 limestones, Washington County; Sample 7: unnamed limestone in Washington Formation, Greene County.

Methods

The core samples were prepared and analyzed by Dr. Rodger E. Denison of the University of Texas at Dallas. The limestones samples were crushed and dissolved in 1 N acetic or HNO_3 acid and insoluable residue was retained in filter paper. Strontium was speciated from other elements using conventional ion-exchange columns. The samples were loaded onto tungsten filaments and analyzed using the procedure outlined in Burke and Hetherington (1988) and Denison et al. (1994). Precision was monitored by routine analysis of the NBS SRM 987 celestite ($SrCO_3$) standard for which a ratio of 0.71014 has been reported.

Results

The $^{87}Sr/^{86}Sr$ ratios of the freshwater limestones range from a low of 0.70968 in the middle Pennsylvanian to 0.71108 for the Upper Pennsylvanian (Table 1). The stratigraphically lowest sample is the Johnstown limestone, which has the second lowest radiogenic $^{87}Sr/^{86}Sr$ ratio, next to the Upper Freeport (?) sample from Broad Top basin, Pennsylvania.

Analyses are also reported with respect to modern seawater, using the delta notation of Burke et al. (1982):

$$\Delta_{sw}=[(^{87}Sr/^{86}Sr)_{unknown} - (^{87}Sr/^{86}Sr)_{modern\ seawater}]x\ 10^5.$$

The weighted mean of modern seawater $^{87}Sr/^{86}Sr$=0.709073b+/- 0.000003 is based on >100 measurements of modern marine shells (Denison et al., 1994).

Discussion

The $^{87}Sr/^{86}Sr$ ratio in the Pennsylvanian-Permian lacustrine limestones from the northern Appalachian basin shows a very slight enrichment through time; measured ratios

Sample	Sample wt (mg)	Insoluble (mg)	net soluble (mg)	% insoluble	$^{87}Sr/^{86}Sr$	Δ_{sw}
unnamed Washington	65.5	12.2	53.3	18.6	0.710635±9	156.2
Benwood #2	62.5	32.7	29.8	52.3	0.710615±15	154.2
Benwood #1	59.3	37.0	22.3	62.4	0.711077±10	200.4
Lower Pittsburgh	57.8	22.6	35.2	39.1	0.710524±16	145.1
Upper Freeport-Outcrop	58.5	16.5	42.0	28.2	0.710374±14	130.1
Upper Freeport-Core	70.3	14.7	55.6	20.9	0.710177±11	110.4
Upper Freeport (?)	53.5	32.0	21.5	59.8	0.709676±15	60.3
Johnstown	70.0	4.4	65.6	6.3	0.709880±13	80.7

Table 5.1. Results of the $^{87}Sr/^{86}Sr$ analyses on freshwater limestones from the Appalachian basin. The Upper Freeport (?) sample is from Broad Top basin in central Pennsylvania. The Upper Freeport-Core sample is from the core collected at Craynes Run in Greene County, Pennsylvania. The Upper Freeport-Outcrop sample was collected at an outcrop in Gallitzen, Pennsylvania. The values of $^{87}Sr/^{86}Sr$ show a progressive trend to higher values through the Pennsylvanian and Permian. The table is organized with the stratigraphically youngest sample at the top, getting progressively older toward the bottom.

range from 0.70968 in the lower-middle Pennsylvanian limestones to 0.71085 in the lower Permian limestones. These values are significantly higher than marine seawater values in the Pennsylvanian and Permian (Denison et al., 1994). In addition, the global marine $^{87}Sr/^{86}Sr$ ratios decrease from 0.70830 in the middle Pennsylvanian to 0.70793 in the early Permian (Denison et al., 1994).

The Upper Freeport (?) limestone in Broad Top basin possessed the lowest $^{87}Sr/^{86}Sr$ ratio of all of the lacustrine limestones analyzed. Because of the uncertainty of the depositional age of the Upper Freeport (?) limestone in Broad Top basin, it could be stratigraphically lower than presumed, and the low $^{87}Sr/^{86}Sr$ ratios could be indicative of stratigraphic misplacement. Additionally, the Upper Freeport (?) from Broad Top basin is significantly less radiogenic than the two Upper Freeport samples from known stratigraphic position. The Benwood #2 sample has a late diagenetic pore-filling sparite, that may have a $^{87}Sr/^{86}Sr$ ratio different from the Pennsylvanian-age lake waters in which the primary calcite was precipitated; therefore, the ratio of the Benwood #2 sample is unreliable. The $^{87}Sr/^{86}Sr$ ratio of the Benwood #1 limestone is probably a more reliable estimate of the geochemical composition of the waters draining the Alleghanian orogen.

All of the limestone samples analyzed for $^{87}Sr/^{86}Sr$ were also analyzed for $\delta^{18}O$ and $\delta^{13}C$ isotopic compositions (Chapter 4, Table 1). High resolution analysis of the freshwater limestones for $\delta^{18}O$ and $\delta^{13}C$ suggests that the samples were deposited in an evaporative setting (Chapter 4). Because the $^{87}Sr/^{86}Sr$ analysis was done on a bulk sample of the limestone, it is necessary to ensure that the mean stable isotopic composition of the limestone is similar to the results obtained from higher resolution sampling. Average values of bulk samples (Chapter 4, Table 1) and the higher resolution sampling of the same unit (Chapter 4, Table 2) are in general agreement, suggesting that the bulk $^{87}Sr/^{86}Sr$ ratio should be a good average for the geochemistry of the Pennsylvanian-Permian freshwater limestones.

The global Pennsylvanian and early Permian marine $^{87}Sr/^{86}Sr$ ratios are much lower than our measured range of the freshwater limestones (Table 2). Marine waters may have ~50 times as much strontium as freshwater, and will alter the freshwater $^{87}Sr/^{86}Sr$ ratio at salinities as low as 10 ppt (Bryant et al., 1995). Thus, there is very little possibility that the observed geochemical properties of the freshwater limestones within the Pennsylvanian and Permian deposits of the Appalachian basin result from mixing of marine waters and freshwater.

In transport-limited (very humid) weathering environments, cation water chemistry reflects the overall cationic composition of the bedrock in the drainage system; whereas in weathering-limited (arid) areas, only the unstable mineral phase cations (commonly Na and Ca) are likely to be dissolved into the water (Stallard, 1992). Stable isotopic data from freshwater limestones indicate that the climate throughout the Pennsylvanian and Permian was periodically arid, but the system probably should be considered as transport-limited, because of the abundance of soil horizons formed by extensive mineral dissolution and plant life that requires meteoric water. The lack of evaporite deposits also suggests a transport-limited system. Therefore, the $^{87}Sr/^{86}Sr$ ratios obtained for the freshwater limestones probably reflect the overall composition of the bedrock in the drainage system.

The subtle rise in the $^{87}Sr/^{86}Sr$ ratios of the freshwater limestones through time reflects a progressive change to more radiogenic (older) sources in the drainage network.

Sample	$^{87}Sr/^{86}Sr$	Depositional Age	$^{87}Sr/^{86}Sr_{paleo-sw}$	$\Delta_{paleo-sw}$
Washington	0.710635	Early Wolfcampian	0.70793	270.5
Benwood #2	0.710615	Late Virgilian	0.70809	252.5
Benwood #1	0.711077	Late Virgilian	0.70809	298.7
Lower Pittsburgh	0.710524	Early Virgilian	0.7081	242.4
Upper Freeport-Outcrop	0.710374	Desmoinian	0.7083	207.4
Upper Freeport-Core	0.710177	Desmoinian	0.7083	187.7
Upper Freeport-? BTB	0.709676	Desmoinian	0.7083	137.6
Johnstown	0.70988	Desmoinian	0.7083	158

Table 5.2. Comparison of analyzed $^{87}Sr/^{86}Sr$ values for the Pennsylvanian-Permian limestones compared to the value of global seawater at the time of deposition (values from Denison et al., 1994). Delta notation is the same as Burke et al. (1982), except that the paleoseawater value is used instead of the modern reference point. All of the freshwater limestones are substantially enriched in radiogenic strontium, indicating that they are of freshwater origin.

Detrital zircons from the Pennsylvanian and Permian sandstones in the Appalachian basin also show a systematic shift in the detrital-zircon-age populations (see Chapter 3). The mineralogical composition of framework grains in lower to middle Pennsylvanian sandstones also reflects exposure of a feldspathic source (Davis and Ehrlich, 1974; O'Connor, 1989). The Pennsylvanian deposits have a substantial population (~14% of the total) of Archean and 1600-1900 Ma detrital zircons that are not represented in the Permian deposits. Instead, Permian sandstones are dominated by Grenville (900-1300 Ma) age zircons. The Archean and 1600-1900 Ma detrital-zircon populations are interpreted to have been recycled from the late Proterozoic/early Paleozoic Laurentian synrift and passive-margin sandstones that were exhumed during the early phases of the Alleghanian orogeny (Thomas et al., 2004a). The lack of these older populations in the Permian sandstones suggests that the exposed orogen provenance did not include the synrift and passive-massive margin deposits. Because the late Proterozoic/early Paleozoic Laurentian passive-margin strata are mostly quartz arenites (e.g. Walker et al., 1994), and do not have minerals that are likely to have contributed radiogenic strontium, the paired elevation of $^{87}Sr/^{86}Sr$ and change in the detrital-zircon-age population indicates incorporation of a primary (igneous and/or metamorphic) source rock into the Alleghanian hinterland.

A simple mass balance approach was used to resolve the contribution of radiogenic strontium from the weathering of silicates to the total $^{87}Sr/^{86}Sr$ ratio for the late Paleozoic freshwater limestones. The Cambrian-Ordovician carbonate shelf is the thickest and most extensive carbonate sequence within the Appalachian thrust belt. For the purpose of this mass balance approach, it was assumed that most of the carbonate exposed in the Pennsylvanian-Permian drainage networks was also of Cambrian-Ordovician age with an approximate average $^{87}Sr/^{86}Sr$ ratio of 0.7090 (Denison et al., 1998; Veizer et al., 1999). It was also assumed that the carbonates contributed 65-85% of the total dissolved Sr in the ancient river waters (Brass, 1976). The calculated $^{87}Sr/^{86}Sr$ ratios of the silicate rocks of the Alleghanian orogen are listed in Table 3. The calculated $^{87}Sr/^{86}Sr$ ratios range from 0.7109 to 0.7228, depending on what proportion of the dissolved strontium is derived from carbonate (65-85%, Table 3), and what proportion is from silicate weathering (15-35%).

For comparison, Albarede and Michard (1987) attempted to evaluate the unroofing history of the Pyrenean orogen by analyzing the $^{87}Sr/^{86}Sr$ ratios of a 65 million-year record of freshwater limestones that were deposited coevally with the exhumation and exposure of crystalline Hercynian basement. The ratios range from 0.7073 and 0.7084 at 55-65 Ma to 0.7077-0.7096 since 30 Ma. Instead of the expected dramatic increase in $^{87}Sr/^{86}Sr$ ratios of the limestones as the pre-Hercynian crystalline basement was exposed and denuded, the ratio remained essentially invariant through time. Although rivers draining the hinterland of the Pyrenees have higher $^{87}Sr/^{86}Sr$ (0.7130-0.7106), the concentration of Sr is quite low and is swamped by the comparatively high concentration of Sr that is supplied from less radiogenic Mesozoic marine sedimentary deposits ($^{87}Sr/^{86}Sr = \sim 0.7075$) that cover the crystalline basement. The data from the Pyrenees suggest that very subtle changes in the $^{87}Sr/^{86}Sr$ ratios in the freshwater limestones may signify important changes in the composition of crust exposed in the drainage network.

Sample	$^{87}Sr/^{86}Sr$	Depositional Age	$^{87}Sr/^{86}Sr$ silicate contribution 65%	75%	85%
Washington	0.710635	Early Wolfcampian	0.7137	0.7155	0.7199
Benwood #2	0.710615	Late Virgilian	0.7136	0.7155	0.7198
Benwood #1	0.711077	Late Virgilian	0.7149	0.7173	0.7228
Lower Pittsburgh	0.710524	Early Virgilian	0.7134	0.7151	0.7192
Upper Freeport-Outcrop	0.710374	Desmoinian	0.7129	0.7145	0.7182
Upper Freeport-Core	0.710177	Desmoinian	0.7124	0.7137	0.7168
Upper Freeport-? BTB	0.709676	Desmoinian	0.7109	0.7117	0.7135
Johnstown	0.70988	Desmoinian	0.7115	0.7125	0.7149

Table 5.3 Estimation of $^{87}Sr/^{86}Sr$ of the silicate fraction of the Alleghanian orogen. Values were calculated assuming that the carbonate contribution to the orogen was 0.709, which is a rough average value for the Cambrian-Ordovician. The Cambrian-Ordovician carbonate platform is the thickest and most extensive carbonate units along the Appalachian margin. Values were calculated assuming that the dissolved $^{87}Sr/^{86}Sr$ was composed of 65%, 75%, and 85% carbonate.

The trend of decreasing $^{87}Sr/^{86}Sr$ global seawater ratios (Denison et al., 1994) over the same time interval that lacustrine $^{87}Sr/^{86}Sr$ values increased also suggests that diagenetic reactions and hydrothermal processes at mid-ocean ridges, including low-temperature alteration of basalt, can counterbalance most increases from continental runoff (Spooner, 1976; Palmer and Edmond, 1989; Richter et al., 1992). The late Paleozoic opening of the Paleotethys Ocean, and the associated hydrothermal alteration of mid-ocean ridge basalt, in eastern and southern Eurasia was probably the most important factor in driving down marine $^{87}Sr/^{86}Sr$ values in the Permian.

Conclusions

The $^{87}Sr/^{86}Sr$ ratios of the Pennsylvanian-Permian freshwater limestones are in excess of marine values, indicating that limestone deposition was free of marine influence. This counters a suggestion by Cassle et al. (2003) that the presence of Spirorbis in freshwater limestones is indicative of deposition in a brackish-water environment open to exchange with marine waters. The $^{87}Sr/^{86}Sr$ values show a trend of slightly increasing values through time, which is interpreted to reflect weathering of silicate mineral phases with more radiogenic $^{87}Sr/^{86}Sr$ in the Appalachian drainage network through time. This roughly mirrors observations of the detrital-zircon population from the Pennsylvanian-Permian succession (Chapter 3), which suggests that clastic sediment was being derived from crystalline basement. If the radiogenic strontium values from freshwater limestones in the northern Appalachian basin are representative of runoff along the entire central and southern Appalachian Mountain belt during the Pennsylvanian-Permian, then it did not significantly affect the isotopic composition of seawater strontium, despite a location in a seasonally wet transport-limited environment at tropical latitudes.

CHAPTER 6: SUMMARY; THE COMPOSITION OF THE ALLEGHANIAN OROGEN

The purpose of this dissertation is to add additional constrains about the tectonic history of the late Paleozoic Alleghanian orogeny by application of isotopic sedimentary provenance proxies to the Pennsylvanian and Permian deposits preserved within the Appalachian and Black Warrior basins. Over the past 15 years, our knowledge of the tectonic evolution of the Alleghanian orogen has grown because of several studies focused in the crystalline hinterland. These studies have demonstrated that the Alleghanian orogeny did not develop solely as the result of subduction of oceanic lithosphere between the Gondwanan and Laurentian continents. Rather, the orogen likely resulted from transpressional deformation along the margin of Laurentia. A geological model accounting for Alleghanian deformation that incorporates the late Paleozoic south-facing Ouachita trench (Keller et al., 1989) remains as a challenge to future geologists.

In response to the uplift of the late Paleozoic Alleghanian orogen, the Appalachian basin filled with erosional detritus (Thomas, 1977). The Black Warrior basin also received sediment from the late Paleozoic Alleghanian orogen and the Ouachita orogen as well (Thomas, 1974, 1977; Whiting and Thomas, 1994; Mars and Thomas, 1999). Petrographic studies of the late Paleozoic deposits in these basins have been used in attempts to reconstruct the tectonic assembly of the orogen (e.g., Davis and Ehrlich, 1974; Graham et al., 1976, Mack et al., 1983; Dickinson et al., 1983; O'Connor, 1989; Osborne, 1991). Many of these studies relied on the premise that the Alleghanian orogen developed in response to a subduction margin collision that resulted in the propagation of thrusts into the Appalachian foreland, and that the orogenic edifice was composed of a diverse mixture of thrust sheets of lower Paleozoic rocks and a volcanic arc terrane. The conclusions to be drawn from the petrographic studies of the Pennsylvanian-Permian deposits in the Appalachian basin have the potential to contaminate studies of other foreland basins because of the improper inference of the Laurentian tectonic environment during the late Paleozoic. In light of our growing understanding of the tectonic environment in which the Alleghanian hinterland was exposed, and to provide context for the late Paleozoic sedimentary record in the Appalachian basin, it is critical to develop a chronological tie between the style of Laurentian plate deformation and sedimentary deposition.

The detrital-zircon-age population of Pennsylvanian and Permian deposits in the Appalachian and Black Warrior basins has succeeded in creating a lithologic "profile" or constituency of the late Paleozoic Alleghanian orogen. The early Pennsylvanian deposits have detrital-zircon-age populations that indicate derivation from recycled Laurentian crust (Thomas et al., 2004a; Chapter 2), and likely included a substantial contribution from upper Proterozoic and lower Paleozoic sedimentary deposits. Early Permian deposits have similar detrital-zircon-age populations, with the exception that they lack Paleoproterozoic- and Archean-age detrital zircons. This is interpreted to reflect a change in the source region, perhaps related to the erosional unroofing of the upper Proterozoic and lower Paleozoic sedimentary section and exposure of the metasedimentary and igneous basement. It is suggested in Chapter 3 that the early Permian deposits were derived from exhumation of the Blue Ridge province or Goochland terrane associated with northwestward-vergent shortening. Both regions are composed of crust with zircon-age populations (Bream, 2002; Owens and Tucker, 2003) similar to those in the Permian deposits (Chapter 3).

An additional constraint on the evolution of the Alleghanian orogeny can be established by analysis of detrital thermochronometers. White mica (muscovite and phengite) is an excellent thermochronometer because of its resilience during weathering and transport and its predictable argon diffusion characteristics. The $^{40}Ar/^{39}Ar$ or K/Ar age of white mica represents the time since the mineral cooled below approximately 350°C (McDougall and Harrison, 1999). K/Ar ages of bulk detrital white mica samples from the Pennsylvanian and Permian sandstones of the Appalachian basin reveal that the source rock cooled below 350°C during the middle Devonian (Aronson and Lewis, 1994; Chapter 3). This age corresponds to the timing of the Acadian orogeny along the eastern Laurentian margin. This result implies that the rocks comprising the Alleghanian orogenic edifice were exhumed from relatively shallow levels throughout the early Permian (<14 km, assuming a geothermal gradient of 25°C/km). The lack of synorogenic thermochronologic ages in the sedimentary record hints that lateral translation of crust may have been an important tectonic process (Willet et al., 2003).

In addition to detrital-zircon-age populations and detrital-white-mica K/Ar ages, $^{87}Sr/^{86}Sr$ was also analyzed from middle Pennsylvanian-Permian lacustrine limestones from the northern Appalachian basin for the purpose of characterizing the crustal constituency of the Alleghanian orogen. The $^{87}Sr/^{86}Sr$ values from the limestones demonstrated that they were deposited in an environment isolated from the influence of marine waters, because the $^{87}Sr/^{86}Sr$ ratios are highly enriched compared to Pennsylvanian-Permian global seawater (Denison et al., 1994). The $^{87}Sr/^{86}Sr$ values of the lacustrine limestones also show an increase from the Pennsylvanian to the Permian. The increase is most likely from silicate mineral dissolution in the drainage network. This indicates that the Alleghanian orogenic edifice is composed either of older, more radiogenic crust or of igneous and metamorphic rocks with mineralogic constituents that are enriched in radiogenic strontium. Sedimentary petrography and detrital-zircon ages from Pennsylvanian and Permian deposits suggest that this upward increase in radiogenic strontium reflects exposure of igneous and metamorphic rocks with labile, ^{87}Sr-rich phases.

An additional phase of research focused on the paleoenvironmental conditions during the deposition of the Pennsylvanian and Permian clastic wedge in the Appalachian basin. Paleobotanical data indicate a substantial shift in Pennsylvanian time from a wet, humid environment to a dry, cool one. The appearance of lacustrine limestones in the Appalachian basin accompanied the shift in plant assemblages, and it has been hypothesized that these changes were results of plate drift to drier latitudes (Cecil, 1990) or development of Appalachian topography that created a “rain shadow” (Rowley et al., 1985; Parrish and Peterson, 1988). The Appalachians are often considered to have been an Andean- to Himalayan-scale orogenic belt (e.g., Slingerland and Furlong, 1989). Paleoclimatic models of the late Paleozoic often assume that the Appalachians would have been sufficiently high to change global atmospheric circulation patterns (e.g., Rowley et al., 1985; Parrish and Peterson, 1988; Kutzbach and Gallimore, 1989; Parrish, 1993; Tabor and Montanez, 2002). Stable isotopic analysis of these limestones revealed that most were deposited in a seasonally wet/dry environment. The water in equilibrium with the lacustrine carbonate was not $\delta^{18}O$ depleted, as might be expected if the

Appalachian basin were in the rain shadow of the Appalachian highlands. Instead, the moisture probably came from epieric seas in the continental interior in a monsoon-like atmospheric circulation system (e.g., Tabor and Montanez, 2002).

Implications of Research

Perhaps the most significant question that rises from this dissertation is: how can geologists infer ancient plate kinematics? In the Appalachians, the generally westward vergence of the Appalachian thrust belt radiating from an igneous and metamorphic hinterland lends to an interpretation of orthogonal collision between Laurentia and Gondwana in the late Paleozoic (e.g., Hatcher and Williams, 1982). This observation was part of the basis for interpreting the Pennsylvanian-Permian clastic deposits as a record of the collision between Gondwana and Laurentia. As this dissertation has pointed out, structural geologists have noted that lateral translation was the dominant displacement direction in the Alleghanian hinterland. In addition, there is very little evidence that Gondwanan crust was incorporated into the Pennsylvanian and Permian sedimentary deposits in the Appalachian basin.

A modern example of the disparity between structural vergence and relative plate motion can be found in southern California. A series of east-vergent compressional structures, including the Coast Ranges, is presently growing despite the northward translation of the Pacific plate relative to the North America plate (Zoback et al., 1987). In addition, the maximum principle stress in regions surrounding the San Andreas fault is perpendicular to fault displacement, which explains the east-west vergence of compressional structures (Zoback et al., 1987). These observations have profound implications for geologists working on plate-scale paleogeographic reconstructions.

How would our understanding of the Alleghanian orogeny change if the tectonic environment were similar to that of modern California? Acknowledgement of these potential complications to our interpretations of the tectonic history of the Alleghanian orogeny is not disparaging to past observations. Instead, it encourages future research initiatives. Some examples of potential lines of research include the following.

1) Continue to build the relationship between detrital-zircon population and petrography of framework grains in the Appalachian basin. Clearly, traditional petrography cannot resolve potentially significant changes in the source of sediment. At what level in the stratigraphic record is the last of the supply of Archean- and Paleoproterozoic-age detrital zircons? Does this correspond to an increase in the proportion of feldspar in the deposits? How does the detrital-zircon-age population from the northern Appalachian-derived Sharon/Olean Conglomerate (Meckel, 1967) and Sharp Mountain Member of the Pottsville Formation compare with detrital-zircon-age populations from southern Appalachian sources? What does the Mississippian Mauch Chunk Formation/Group represent in the overall tectonic history of the Alleghanian orogeny?
2) Combine $^{40}Ar/^{39}Ar$ dating of detrital white mica with petrography and detrital-zircon-age populations to try to match the tectonic history of the Alleghanian hinterland with the sedimentological record in the Appalachian basin. Present technology allows for inexpensive, single-grain $^{40}Ar/^{39}Ar$ analysis of detrital

white mica and feldspar, which can be applied to the Mississippian, Pennsylvanian, and Permian sandstones from the Appalachian and Black Warrior basins. Detrital-zircon-age populations can be used to determine the source of the sediment, while thermochronometry can inform us of the rate at which the source rock got to the surface. Are there along-strike variations in the ages of white mica in the Appalachian and Black Warrior basins?

3) Expand the initial isotopic study of lacustrine limestones for the purpose of relating the data to paleobotanical studies. Do additional stable isotopic analyses of the limestones result in a consistent interpretation of periodic wet/dry conditions in the Appalachian basin? Are the $^{87}Sr/^{86}Sr$ values of the limestones consistent along strike? Is the upward enrichment in $^{87}Sr/^{86}Sr$ in the limestones a robust trend throughout the Appalachian basin?

The research in this dissertation cannot definitively be used to defend a single, unified theory of Alleghanian orogenesis. Instead, it provides a basis and template for future research endeavors that may help bring resolution to the many problems in existing models regarding the evolution of the Appalachians. Paradoxically, the complexity of this geologic problem can only be reduced to a simpler solution by the acquisition of more data.

Appendix A. U-Pb geochronologic analyses by Laser-Ablation Multicollector ICP Mass Spectrometry

	Isotopic ratios						Apparent ages (Ma)					
U (ppm)	$^{206}Pb/^{204}Pb$	$^{207}Pb^*/^{235}U$	± (%)	$^{206}Pb^*/^{238}U$	± (%)	error corr.	$^{206}Pb^*/^{238}U$	± (Ma)	$^{207}Pb^*/^{235}U$	± (Ma)	$^{206}Pb^*/^{207}Pb^*$	± (Ma)
Tumbling Run Member, Pottsville Formation of eastern Pennsylvania												
92	20.268	0.4232	2.7307	0.0622	1.08	0.4	*389.0*	*4.3*	358.3	11.7	163.9	29.3
20	18.322	0.4910	8.2639	0.0652	1.19	0.1	*407.5*	*5.0*	405.6	40.4	395.1	91.7
50	20.774	0.4420	7.9125	0.0666	1.39	0.2	*415.6*	*6.0*	371.7	34.9	106.0	92.0
51	18.664	0.4956	3.0886	0.0671	0.55	0.2	*418.6*	*2.4*	408.7	15.4	353.4	34.3
24	21.703	0.4283	9.3030	0.0674	0.98	0.1	*420.6*	*4.3*	362.0	39.7	1.6	111.5
18	19.032	0.4889	11.4762	0.0675	1.01	0.1	*421.0*	*4.4*	404.1	55.4	309.1	130.1
75	19.392	0.4891	2.3411	0.0688	1.01	0.4	*428.8*	*4.5*	404.3	11.6	266.3	24.2
42	17.119	0.5572	2.4859	0.0692	0.70	0.3	*431.2*	*3.1*	449.7	14.0	545.3	26.1
57	12.275	0.7946	3.4690	0.0707	1.00	0.3	*440.6*	*4.6*	593.8	27.6	1232.7	32.6
37	18.020	0.5862	3.3328	0.0766	0.82	0.2	*475.8*	*4.0*	468.4	19.6	432.2	36.0
17	18.293	0.6353	6.0874	0.0843	1.44	0.2	*521.7*	*7.8*	499.4	38.5	398.6	66.3
14	16.307	0.7602	10.4660	0.0899	1.41	0.1	*555.0*	*8.1*	574.1	77.7	650.6	111.3
35	13.337	1.5764	1.8708	0.1525	1.13	0.6	*914.9*	*11.1*	961.0	29.5	1067.9	15.0
11	13.403	1.7059	4.8306	0.1658	1.51	0.3	*989.1*	*16.1*	1010.8	80.4	1058.0	46.2
76	13.091	1.9145	0.7714	0.1818	0.62	0.8	*1076.7*	*7.2*	1086.2	14.9	1105.2	4.6
4	14.813	1.7118	8.6927	0.1839	1.17	0.1	*1088.3*	*13.9*	1012.9	140.8	853.5	89.5
83	12.476	2.2038	0.9857	0.1994	0.88	0.9	1172.2	11.3	1182.3	21.8	*1200.8*	4.4
10	13.313	2.1039	2.8341	0.2031	1.20	0.4	*1192.1*	15.7	1150.1	58.8	1071.6	25.8
13	12.036	2.3909	2.6133	0.2087	0.78	0.3	1221.9	10.5	1239.9	61.5	*1271.2*	24.3
18	12.228	2.3929	2.3730	0.2122	0.59	0.2	1240.6	8.1	1240.5	56.1	*1240.2*	22.5
26	11.181	2.7719	1.7359	0.2248	1.00	0.6	1307.1	14.5	1348.0	47.7	*1413.6*	13.5
31	11.325	2.8735	1.3621	0.2360	0.92	0.7	1365.9	14.0	1375.0	39.0	*1389.1*	9.6
44	10.695	3.1431	1.3411	0.2438	0.98	0.7	1406.4	15.3	1443.3	41.9	*1498.1*	8.7
10	10.714	3.2201	3.8960	0.2502	1.70	0.4	1439.6	27.4	1462.0	120.0	*1494.7*	33.2
27	10.850	3.3802	1.5427	0.2660	1.10	0.7	1520.5	18.9	1499.8	51.6	*1470.7*	10.2
43	5.859	9.5228	1.0062	0.4047	0.99	1.0	2190.4	25.7	2389.7	92.9	*2564.3*	1.7
Pottsville Formation, Broad Top basin of central Pennsylvania												
19	2729.034	0.4498	8.4338	0.0623	1.53	0.2	389.3	6.1	377.2	37.8	303.3	94.5
21	22144.13	0.4867	9.3745	0.0639	1.07	0.1	399.6	4.4	402.6	45.3	420.3	104.0
87	53185.52	0.4905	1.6887	0.0651	0.60	0.4	406.7	2.5	405.3	8.4	396.8	17.7
21	100122.4	0.4386	7.4665	0.0672	0.77	0.1	419.4	3.3	369.3	32.7	65.7	88.4
15	2028	0.5320	12.9687	0.0677	0.81	0.1	422.5	3.6	433.2	67.7	490.3	142.7
34	32325.71	0.5803	4.3878	0.0681	0.80	0.2	424.9	3.5	464.6	25.5	665.9	46.2
15	667.0777	0.6683	10.1227	0.0683	1.03	0.1	426.1	4.5	519.7	66.5	955.2	103.0
109	2088	0.5769	1.9513	0.0699	1.29	0.7	435.6	5.8	462.5	11.4	598.4	15.9
70	3006	0.5207	3.2811	0.0709	0.97	0.3	441.9	4.5	425.6	17.2	338.4	35.5
52	16713.67	0.5450	2.5420	0.0719	0.53	0.2	447.3	2.5	441.7	14.0	412.7	27.8
75	211100.8	0.5672	1.2698	0.0724	0.62	0.5	450.4	2.9	456.2	7.3	485.5	12.2
27	303256.7	0.5430	5.5338	0.0724	0.70	0.1	450.7	3.3	440.4	30.1	387.2	61.6
49	80985.44	0.5902	4.2843	0.0735	0.39	0.1	457.4	1.9	471.0	25.4	537.8	46.7
18	2301	0.5588	10.3069	0.0739	1.18	0.1	459.6	5.6	450.8	56.9	406.0	114.6
79	399702	0.7003	1.1958	0.0887	0.57	0.5	547.9	3.3	539.0	8.5	501.7	11.6
9	2193	1.0023	8.1202	0.1089	3.37	0.4	666.4	23.6	705.0	79.5	829.9	77.0
104	1724.624	1.4765	1.0053	0.1451	0.82	0.8	873.7	7.7	920.8	15.0	1035.4	5.9

Appendix A (cont.). U-Pb geochronologic analyses by Laser-Ablation Multicollector ICP Mass Spectrometry

		Isotopic ratios					Apparent ages (Ma)					
U (ppm)	$^{206}Pb/^{204}Pb$	$^{207}Pb^*/^{235}U$	± (%)	$^{206}Pb^*/^{238}U$	± (%)	error corr.	$^{206}Pb^*/^{238}U$	± (Ma)	$^{207}Pb^*/^{235}U$	± (Ma)	$^{206}Pb^*/^{207}Pb^*$	± (Ma)
Pottsville Formation, Broad Top basin of central Pennsylvania (cont.)												
108	5434	1.5519	0.9623	0.1454	0.57	0.6	875.0	5.3	951.2	15.1	1132.1	7.7
9	4404	1.4968	7.8517	0.1511	0.83	0.1	906.9	8.1	929.1	112.8	982.1	79.5
22	3535.137	1.4309	2.5572	0.1579	0.73	0.3	945.2	7.4	901.9	36.5	797.4	25.7
30	6251.058	1.6660	1.8219	0.1603	0.58	0.3	958.5	6.0	995.7	30.4	1078.4	17.3
23	23267.5	1.5790	2.7050	0.1658	1.06	0.4	988.8	11.3	962.0	42.5	901.1	25.7
7	2723.335	1.6258	6.3807	0.1659	0.96	0.1	989.5	10.2	980.2	100.2	959.6	64.5
17	2455.457	1.6322	4.2391	0.1680	1.19	0.3	1001.3	12.8	982.7	67.9	941.4	41.7
7	1034.096	1.7321	7.3932	0.1693	1.42	0.2	1008.3	15.4	1020.5	122.4	1046.9	73.2
27	16943.02	1.7912	2.3294	0.1707	0.75	0.3	1016.2	8.3	1042.3	41.5	1097.3	22.1
104	5586	1.7218	0.9490	0.1735	0.58	0.6	1031.2	6.5	1016.7	16.5	985.7	7.6
12	1962.724	1.4741	4.2537	0.1746	0.87	0.2	1037.3	9.8	919.8	61.8	647.6	44.7
11	2744.914	1.7308	4.6768	0.1758	0.97	0.2	1043.8	11.0	1020.1	79.0	969.4	46.7
13	15454.9	1.9176	3.4534	0.1758	0.99	0.3	1044.1	11.2	1087.2	65.1	1174.6	32.7
16	19082.43	1.9456	3.6336	0.1759	1.11	0.3	1044.3	12.5	1096.9	69.4	1203.0	34.1
4	4073	2.0666	11.2492	0.1762	2.23	0.2	1046.3	25.2	1137.8	212.2	1316.7	106.9
128	5804.624	1.8299	1.0833	0.1800	0.88	0.8	1067.1	10.2	1056.3	19.9	1034.0	6.4
70	21900	1.9239	1.2069	0.1809	0.74	0.6	1071.8	8.6	1089.4	23.3	1125.0	9.5
18	61029.75	1.9408	2.5127	0.1815	0.77	0.3	1075.2	9.0	1095.3	48.3	1135.4	23.8
26	53678.5	1.9194	1.8886	0.1820	0.44	0.2	1077.6	5.2	1087.8	36.2	1108.3	18.3
17	13207.95	1.8644	2.3788	0.1831	0.54	0.2	1084.0	6.3	1068.6	44.1	1037.2	23.4
193	9934	1.8688	0.8523	0.1836	0.69	0.8	1086.4	8.1	1070.1	16.0	1037.0	5.1
11	2598.536	1.8143	4.6939	0.1840	0.75	0.2	1089.0	8.9	1050.6	83.0	971.7	47.3
64	57136	1.9605	0.7542	0.1842	0.39	0.5	1089.6	4.6	1102.1	14.9	1126.7	6.5
8	5332.37	2.0033	8.5117	0.1853	1.12	0.1	1096.0	13.3	1116.6	159.9	1157.0	83.7
11	2591.468	1.8158	2.7500	0.1853	0.68	0.2	1096.1	8.2	1051.2	49.5	959.0	27.2
135	1011095	1.9610	0.8610	0.1855	0.78	0.9	1096.7	9.3	1102.2	17.0	1113.1	3.7
62	9704.868	1.9788	1.0605	0.1885	0.79	0.7	1113.2	9.6	1108.3	21.1	1098.7	7.0
28	43645.31	2.1276	1.5857	0.1885	0.60	0.4	1113.3	7.3	1157.8	33.7	1242.2	14.4
26	33081.28	2.1605	1.4189	0.1890	0.83	0.6	1115.8	10.1	1168.4	30.7	1267.4	11.2
7	1951	2.0987	5.1683	0.1914	0.90	0.2	1129.2	11.1	1148.4	104.6	1184.9	50.3
8	2557.865	2.1912	5.6605	0.1954	1.58	0.3	1150.7	19.9	1178.3	118.7	1229.2	53.4
42	8437	2.1885	1.3321	0.1957	0.71	0.5	1152.4	8.9	1177.4	29.2	1223.6	11.1
6	14009.69	2.3848	4.8533	0.1963	1.39	0.3	1155.5	17.6	1238.1	111.2	1384.8	44.7
36	77838	2.3194	1.5516	0.1977	0.84	0.5	1163.0	10.7	1218.2	35.9	1317.4	12.6
9	116882	2.5733	6.0708	0.1987	2.15	0.4	1168.2	27.5	1293.1	147.4	*1506.8*	*53.6*
20	21714.57	2.3329	2.1889	0.2026	1.06	0.5	1189.3	13.9	1222.4	50.6	1281.2	18.6
4	5167.89	1.7287	8.3717	0.2032	1.19	0.1	1192.7	15.5	1019.3	137.2	663.3	88.8
16	1641.442	2.1558	1.8788	0.2056	0.94	0.5	1205.2	12.5	1166.9	40.3	1096.5	16.3
10	3540.28	2.2188	3.2327	0.2065	0.75	0.2	1210.1	10.0	1187.0	70.3	1145.1	31.2
67	1680	2.6721	2.2354	0.2082	0.78	0.4	1219.0	10.5	1320.8	58.9	*1489.9*	*19.8*
7	10174	2.4498	3.5028	0.2082	1.27	0.4	1219.1	17.0	1257.4	83.6	*1323.4*	*31.6*
45	21244.68	2.3564	1.0768	0.2089	0.75	0.7	1222.8	10.0	1229.5	25.4	*1241.2*	*7.6*
11	4868.43	2.6125	2.7448	0.2112	0.83	0.3	1235.2	11.3	1304.2	70.3	*1419.5*	*25.0*
16	53320.37	2.3688	2.0873	0.2120	0.66	0.3	1239.2	9.1	1233.2	49.0	1222.8	19.4
25	19226.06	2.5974	1.4555	0.2132	1.02	0.7	1245.6	14.0	1299.9	37.7	*1390.7*	*10.0*

Appendix A (cont.). U-Pb geochronologic analyses by Laser-Ablation Multicollector ICP Mass Spectrometry

	Isotopic ratios						Apparent ages (Ma)					
U (ppm)	$^{206}Pb/^{204}Pb$	$^{207}Pb^{*}/^{235}U$	± (%)	$^{206}Pb^{*}/^{238}U$	± (%)	error corr.	$^{206}Pb^{*}/^{238}U$	± (Ma)	$^{207}Pb^{*}/^{235}U$	± (Ma)	$^{206}Pb^{*}/^{207}Pb^{*}$	± (Ma)
Pottsville Formation, Broad Top basin of central Pennsylvania (cont.)												
10	3138.081	2.5470	3.0721	0.2144	1.16	0.4	1252.1	16.0	1285.6	76.5	*1341.8*	*27.5*
7	27406	2.5187	4.2800	0.2177	1.08	0.3	1269.5	15.1	1277.4	104.0	*1290.8*	*40.3*
23	41583.79	2.7630	2.0416	0.2241	1.03	0.5	1303.3	14.8	1345.6	55.7	*1413.4*	*16.9*
13	3665	2.7707	2.9111	0.2329	1.30	0.4	1349.4	19.4	1347.7	78.8	*1344.9*	*25.2*
11	80004.14	2.8011	3.2520	0.2375	0.68	0.2	1373.5	10.3	1355.8	88.5	*1328.1*	*30.8*
12	27043.04	2.9149	2.6321	0.2420	1.28	0.5	1396.9	20.0	1385.8	75.1	*1368.7*	*22.1*
14	6463.117	3.4659	1.5124	0.2480	0.76	0.5	1427.9	12.2	1519.5	51.9	*1649.5*	*12.1*
31	110816.2	3.2224	1.2755	0.2511	0.45	0.4	1444.2	7.3	1462.6	40.9	*1489.4*	*11.3*
30	84605	3.4535	1.2548	0.2566	0.92	0.7	1472.4	15.2	1516.7	43.1	*1579.1*	*8.0*
15	25992.61	4.6787	1.8626	0.3032	1.20	0.6	1707.4	23.5	1763.4	84.8	*1830.5*	*12.9*
20	7754843	5.4748	1.0012	0.3314	0.73	0.7	1845.4	15.5	1896.7	54.2	*1953.2*	*6.1*
56	23960	5.8604	0.6009	0.3353	0.52	0.9	1863.9	11.3	1955.4	35.1	*2053.7*	*2.6*
47	44054.48	6.0740	0.9580	0.3605	0.76	0.8	1984.4	17.7	1986.5	57.4	*1988.7*	*5.2*
4	4514.057	7.4789	3.7582	0.3798	1.57	0.4	2075.2	38.3	2170.5	251.5	*2261.8*	*29.5*
18	6308.397	10.3731	2.4277	0.4243	2.36	1.0	2280.2	64.2	2468.6	228.1	*2627.7*	*4.8*
Pocahontas Formation, Hurricane Ridge syncline of southern West Virginia												
43	4092	1.4651	1.71	0.1527	1.12	0.7	*915.9*	*11.0*	916.1	25.1	916.6	13.3
69	2363	1.5248	1.22	0.1615	0.92	0.8	*965.2*	*9.6*	940.4	18.7	882.9	8.3
121	26937	1.6142	1.52	0.1617	1.20	0.8	*966.2*	*12.5*	975.8	24.6	997.4	9.4
32	20545	1.5994	5.20	0.1663	5.00	1.0	*991.7*	*53.4*	970.0	81.1	921.1	14.6
8	2435	2.0354	4.48	0.1665	1.56	0.3	993.0	16.8	1127.4	88.7	*1396.5*	*40.3*
76	11491	1.6196	1.40	0.1675	1.21	0.9	*998.1*	*13.1*	977.8	22.8	932.7	7.2
197	2979	1.7716	1.56	0.1680	1.36	0.9	*1001.1*	*14.7*	1035.1	27.7	1107.6	7.7
238	1122	1.8135	1.09	0.1687	0.96	0.9	*1005.2*	*10.5*	1050.3	19.8	1145.5	5.0
154	5917	1.8222	1.62	0.1695	0.82	0.5	*1009.2*	*9.0*	1053.5	29.5	1146.5	13.9
40	237511	1.8836	4.2287	0.1702	1.52	0.4	*1013.0*	*16.6*	1075.3	77.8	1204.1	38.9
99	1368098	1.6864	1.61	0.1712	1.28	0.8	*1018.9*	*14.1*	1003.4	27.2	969.8	9.9
34	4593	1.6553	1.52	0.1733	0.62	0.4	*1030.0*	*6.9*	991.6	25.2	907.5	14.3
127	335178	1.7653	0.85	0.1740	0.61	0.7	*1033.9*	*6.8*	1032.8	15.1	1030.4	6.0
186	20885	1.7562	1.25	0.1785	1.24	1.0	*1058.9*	*14.2*	1029.4	22.1	967.4	2.0
6	6594	2.1126	9.34	0.1788	1.48	0.2	*1060.6*	*17.0*	1152.9	182.8	1331.0	89.2
23	48640	1.7810	2.32	0.1791	1.03	0.4	*1062.2*	*11.9*	1038.5	41.1	989.0	21.1
193	1209963	1.7523	2.04	0.1833	0.98	0.5	*1085.0*	*11.5*	1028.0	35.6	908.8	18.4
69	2391	2.0479	1.72	0.1872	0.79	0.5	*1106.2*	*9.5*	1131.6	35.1	1180.7	15.1
19	7940	1.9959	2.24	0.1890	0.85	0.4	*1115.8*	*10.3*	1114.1	44.5	1110.9	20.7
22	10722	1.9990	2.01	0.1902	1.18	0.6	*1122.7*	*14.4*	1115.2	40.0	1100.6	16.3
168	3028	2.1205	1.31	0.1909	0.51	0.4	*1126.3*	*6.3*	1155.5	27.7	1210.7	11.8
19	3935	2.0454	2.41	0.1913	1.14	0.5	*1128.3*	*14.1*	1130.8	48.9	1135.5	21.1
121	36967	2.0691	0.8582	0.1924	0.71	0.8	*1134.6*	*8.9*	1138.6	17.9	1146.3	4.7
82	2379	2.0768	0.85	0.1925	0.65	0.8	*1134.9*	*8.0*	1141.2	17.8	1153.2	5.5
63	131005	2.1023	1.28	0.1926	0.97	0.8	*1135.6*	*12.0*	1149.5	27.0	1175.9	8.3
99	8412	1.9522	0.95	0.1933	0.67	0.7	*1139.2*	*8.3*	1099.2	18.7	1020.8	6.9
25	3910	2.0417	1.68	0.1934	0.88	0.5	*1139.6*	*10.9*	1129.5	34.3	1110.2	14.3
115	8475	1.9873	1.64	0.1937	1.55	0.9	*1141.2*	*19.3*	1111.2	32.6	1053.1	5.5
28	9203	2.1873	1.45	0.1949	1.11	0.8	*1147.7*	*14.0*	1177.0	31.7	1231.3	9.1

Appendix A (cont.). U-Pb geochronologic analyses by Laser-Ablation Multicollector ICP Mass Spectrometry

	Isotopic ratios						Apparent ages (Ma)					
U (ppm)	$^{206}Pb/^{204}Pb$	$^{207}Pb^{*}/^{235}U$	± (%)	$^{206}Pb^{*}/^{238}U$	± (%)	error corr.	$^{206}Pb^{*}/^{238}U$	± (Ma)	$^{207}Pb^{*}/^{235}U$	± (Ma)	$^{206}Pb^{*}/^{207}Pb^{*}$	± (Ma)
Pocahontas Formation, Hurricane Ridge syncline of southern West Virginia (cont.)												
33	2218	2.0627	1.73	0.1962	0.59	0.3	*1154.8*	*7.4*	1136.5	35.5	1101.7	16.2
14	38015	2.1870	3.20	0.1962	0.92	0.3	*1155.0*	*11.7*	1176.9	68.6	1217.4	30.1
8	1157	2.3206	4.62	0.1979	0.70	0.2	*1164.1*	*8.9*	1218.6	103.4	1316.5	44.2
21	21447	2.3263	2.61	0.2003	1.23	0.5	*1177.2*	*15.9*	1220.3	59.8	1297.5	22.3
48	5626	2.2674	1.58	0.2005	1.35	0.9	*1178.2*	*17.4*	1202.2	35.8	1245.7	8.1
27	281640	2.2965	1.46	0.2036	1.05	0.7	*1194.9*	*13.8*	1211.2	33.6	1240.4	10.0
55	82271	2.3209	1.47	0.2043	0.91	0.6	*1198.6*	*12.0*	1218.7	34.1	1254.5	11.3
60	4700	2.0508	1.99	0.2048	0.94	0.5	*1201.2*	*12.4*	1132.6	40.7	1003.3	17.8
64	2328	2.4274	1.09	0.2	0.9	0.8	1219.4	12.1	1250.7	26.4	*1305.1*	*5.9*
34	1164579	2.3544	1.59	0.2089	1.45	0.9	*1222.7*	*19.5*	1228.9	37.4	1239.7	6.5
30	4117	2.3706	1.75	0.2089	0.83	0.5	1222.8	11.2	1233.8	41.3	*1253.1*	*15.0*
40	3498	2.3370	1.79	0.2	1.0	0.6	*1223.2*	*13.5*	1223.6	41.7	1224.3	14.6
19	14403	2.4062	2.28	0.2	0.6	0.2	1226.2	7.6	1244.5	54.3	*1276.2*	*21.6*
9	19112	2.2866	6.36	0.2113	0.73	0.1	*1235.5*	*9.9*	1208.2	137.8	1159.6	62.6
35	13449	2.4326	1.50	0.2115	0.65	0.4	1237.0	8.9	1252.3	36.4	*1278.7*	*13.2*
219	4178	2.4332	1.56	0.2118	0.84	0.5	1238.6	11.4	1252.5	37.8	*1276.4*	*12.84*
34	619872	2.5102	1.44	0.2129	0.76	0.5	1244.3	10.4	1275.0	36.1	*1327.1*	*11.8*
30	11642	2.5667	1.53	0.2139	1.16	0.8	1249.5	16.0	1291.2	39.1	*1361.2*	*9.6*
35	14899	2.2505	1.44	0.2170	0.90	0.6	*1265.8*	*12.5*	1197.0	32.5	1074.8	11.4
62	95840	2.3943	2.43	0.2172	2.26	0.9	*1267.3*	*31.6*	1240.9	57.5	1195.3	8.8
65	29380	2.6440	1.05	0.2217	0.92	0.9	1290.9	13.1	1313.0	27.7	*1349.3*	*4.8*
52	30248	2.8119	1.05	0.2249	0.83	0.8	1307.6	12.0	1358.7	29.7	*1440.0*	*6.2*
34	3604	2.7101	1.32	0.2254	0.78	0.6	1310.5	11.3	1331.2	35.7	*1364.7*	*10.3*
88	7652	2.5702	1.21	0.2287	0.85	0.7	1327.6	12.5	1292.2	31.1	*1233.9*	*8.44*
87	39002	2.7495	0.77	0.2311	0.50	0.7	*1340.2*	*7.5*	1342.0	21.3	1344.7	5.6
51	43153	2.7368	1.00	0.2328	0.77	0.8	1349.1	11.5	1338.5	27.3	*1321.7*	*6.11*
13	16738	3.0216	3.18	0.2418	1.29	0.4	1396.2	20.0	1413.1	93.0	*1438.7*	*27.7*
21	10784	3.1405	1.63	0.2511	0.92	0.6	1444.1	14.9	1442.7	50.5	*1440.5*	*12.8*
12	158426	2.9725	2.34	0.2528	0.59	0.3	*1452.9*	*9.6*	1400.6	68.4	1321.9	22.0
38	77498	3.2193	0.96	0.2589	0.67	0.7	1484.2	11.2	1461.8	30.9	*1429.4*	*6.57*
20	778184	3.3560	1.67	0.2616	0.99	0.6	1498.3	16.7	1494.2	55.4	*1488.4*	*12.7*
74	12730	7.2568	0.41	0.4016	0.36	0.9	2176.4	9.4	2143.5	29.5	*2112.2*	*1.62*
Lee Formation, Little Stone Gap of western Virginia												
36	23573	1.4376	2.24	0.1493	0.61	0.3	*897.1*	*5.9*	904.7	32.2	923.6	22.2
218.1	8911	1.5948	0.83	0.1514	0.71	0.9	*909.1*	*6.9*	968.2	13.4	1104.9	4.4
9	4366	1.6807	6.44	0.1530	0.84	0.1	*917.5*	*8.3*	1001.2	104.3	1189.4	63.0
24	23122	1.7773	2.51	0.1572	0.78	0.3	*941.2*	*7.9*	1037.2	44.3	1245.6	23.4
34	40590	1.8653	2.13	0.1646	1.67	0.8	*982.5*	*17.7*	1068.9	39.5	1249.5	12.9
9.154	2783	1.7853	5.38	0.1652	1.10	0.2	*985.6*	*11.7*	1040.1	93.1	1156.5	52.2
50	138438	1.7587	1.17	0.1677	0.81	0.7	*999.6*	*8.7*	1030.4	20.8	1096.2	8.5
90.54	165976	1.8407	1.24	0.1724	1.00	0.8	*1025.3*	*11.1*	1060.1	23.0	1132.4	7.3
28.67	29616	1.9186	1.88	0.1752	1.30	0.7	*1040.5*	*14.7*	1087.6	35.9	1183.0	13.3
30	130402	1.8535	1.76	0.1754	0.48	0.3	*1041.6*	*5.5*	1064.7	32.6	1112.2	16.9
7	162183	1.9202	5.25	0.1769	1.28	0.2	*1050.2*	*14.5*	1088.1	97.6	1164.9	50.5
72	96011	1.9252	0.96	0.1778	0.75	0.8	*1055.1*	*8.5*	1089.9	18.6	1160.1	6.0

Appendix A (cont.). U-Pb geochronologic analyses by Laser-Ablation Multicollector ICP Mass Spectrometry

	Isotopic ratios						Apparent ages (Ma)					
U (ppm)	$^{206}Pb/^{204}Pb$	$^{207}Pb^*/^{235}U$	± (%)	$^{206}Pb^*/^{238}U$	± (%)	error corr.	$^{206}Pb^*/^{238}U$	± (Ma)	$^{207}Pb^*/^{235}U$	± (Ma)	$^{206}Pb^*/^{207}Pb^*$	± (Ma)
Lee Formation, Little Stone Gap of western Virginia (cont.)												
71.4	229575	2.1593	1.39	0.1806	1.19	0.9	*1070.5*	*13.8*	1168.0	29.9	1353.7	6.8
17	16483	2.0515	2.80	0.1813	0.94	0.3	*1074.3*	*10.9*	1132.8	56.7	1246.6	25.8
13	2251	2.0435	3.68	0.1850	1.27	0.3	*1094.4*	*15.1*	1130.1	73.7	1199.3	34.1
11	7325	2.3875	5.06	0.1859	1.28	0.3	*1098.9*	*15.4*	1238.9	115.8	1491.3	46.3
20	66886	2.0738	2.70	0.1859	0.81	0.3	*1099.3*	*9.7*	1140.2	55.4	1218.8	25.4
9	6756	1.9641	3.38	0.1859	0.45	0.1	*1099.4*	*5.4*	1103.3	65.3	1111.1	33.5
40	299222	2.2369	2.31	0.1861	1.00	0.4	*1100.0*	*12.0*	1192.7	51.1	1364.9	20.0
16	4260	2.1429	2.98	0.1864	1.41	0.5	*1102.0*	*17.0*	1162.8	62.8	1277.8	25.5
25	135438	2.0664	1.85	0.1873	0.73	0.4	*1107.0*	*8.8*	1137.7	38.0	1196.9	16.7
11	6630	2.2967	7.21	0.1885	2.87	0.4	*1113.3*	*34.8*	1211.3	155.5	1390.4	63.4
44	1601	2.0380	3.26	0.1902	1.22	0.4	*1122.7*	*14.9*	1128.3	65.3	1139.1	30.1
24	12483	2.1402	1.44	0.1912	0.65	0.5	*1128.0*	*8.0*	1161.9	30.8	1225.7	12.6
16	11103	2.0148	3.51	0.1924	0.71	0.2	*1134.4*	*8.8*	1120.5	69.4	1093.8	34.4
35.66	48502	2.3907	1.60	0.2021	0.89	0.6	*1186.7*	*11.5*	1239.8	38.0	1333.4	12.8
12	32026	2.1204	3.08	0.2031	0.74	0.2	*1192.0*	*9.7*	1155.5	64.3	1087.7	30.0
64.32	48975	2.4357	0.91	0.2035	0.61	0.7	*1194.3*	*8.0*	1253.2	22.3	1355.9	6.5
80	6562	2.4032	1.55	0.2087	1.21	0.8	1221.7	16.3	1243.6	37.0	*1281.6*	*9.4*
19.35	25661	2.8744	2.39	0.2175	1.65	0.7	1268.5	23.2	1375.2	67.5	*1545.0*	*16.2*
30	38110	2.8623	1.55	0.2233	1.15	0.7	1299.4	16.6	1372.0	44.0	*1487.0*	*9.8*
51.28	50404	2.7985	1.18	0.2253	0.79	0.7	1309.6	11.5	1355.1	33.0	*1427.8*	*8.3*
60.82	117327	2.8048	0.96	0.2266	0.64	0.7	1316.5	9.3	1356.8	27.1	*1420.8*	*6.9*
10	10851	2.8230	4.75	0.2298	2.08	0.4	1333.4	30.8	1361.7	127.8	*1406.3*	*40.9*
6	8469	2.9407	6.32	0.2343	2.32	0.4	1357.0	34.9	1392.4	173.1	*1447.1*	*56.0*
104.7	6110	3.7149	1.01	0.2546	0.93	0.9	1462.0	15.3	1574.6	37.5	*1728.9*	*3.6*
35.57	381869	3.4573	1.01	0.2552	0.56	0.6	1465.4	9.1	1517.5	34.7	*1591.1*	*7.8*
73.72	23695	3.7745	0.93	0.2651	0.85	0.9	1515.9	14.5	1587.3	35.1	*1683.6*	*3.5*
38.82	4083	3.9841	1.71	0.2701	1.45	0.9	1541.1	25.2	1631.0	66.7	*1748.9*	*8.2*
54.51	30262	4.0289	0.96	0.2713	0.79	0.8	1547.6	13.8	1640.0	38.6	*1760.7*	*5.0*
20	56437	3.7119	1.49	0.2718	0.85	0.6	1550.1	14.8	1573.9	54.8	*1606.0*	*11.5*
38	64636	3.8637	1.07	0.2807	0.64	0.6	1594.9	11.7	1606.1	41.0	*1620.8*	*7.9*
25	19697	4.0046	1.54	0.2865	0.84	0.5	1624.0	15.5	1635.1	60.8	*1649.5*	*12.0*
8	34027	3.5855	5.03	0.2869	2.49	0.5	1625.8	46.0	1546.3	168.3	*1439.3*	*41.6*
68	46255	4.1888	0.75	0.2876	0.58	0.8	1629.7	10.7	1671.8	31.5	*1725.2*	*4.4*
52.19	60759	4.2722	1.02	0.2933	0.79	0.8	1657.9	14.9	1688.0	43.5	*1725.7*	*6.0*
12.23	6975	4.6825	2.59	0.2981	1.02	0.4	1682.0	19.5	1764.1	116.3	*1862.8*	*21.5*
28	17582	4.5909	1.08	0.2986	0.75	0.7	1684.6	14.5	1747.6	48.9	*1823.8*	*7.0*
10	41147	4.3370	2.67	0.3034	1.81	0.7	1708.2	35.4	1700.4	111.2	*1690.9*	*18.0*
76	57775	4.7553	0.85	0.3071	0.80	0.9	1726.3	15.9	1777.0	40.2	*1837.2*	*2.5*
60.36	61154	4.7540	1.02	0.3091	0.72	0.7	1736.1	14.2	1776.8	48.1	*1825.0*	*6.6*
12	189341	6.7835	1.32	0.3707	0.64	0.5	2032.8	15.3	2083.6	86.9	*2134.1*	*10.1*
36.87	14417	10.1627	1.56	0.3992	1.40	0.9	2165.5	36.0	2449.7	149.2	*2694.8*	*5.6*
172.6	19004	10.3864	1.89	0.4186	1.88	1.0	2254.0	50.4	2469.8	181.6	*2652.6*	*1.6*
82.76	33515	13.3640	1.11	0.4565	1.04	0.9	2423.9	30.5	2705.7	140.9	*2923.5*	*3.2*
37	7120	11.9839	1.12	0.4592	0.85	0.8	2435.9	25.2	2603.1	127.8	*2736.0*	*6.0*

Appendix A (cont.). U-Pb geochronologic analyses by Laser-Ablation Multicollector ICP Mass Spectrometry

U	^{206}Pb	Isotopic ratios $^{207}Pb^*$	±	$^{206}Pb^*$	±	error	Apparent ages (Ma) $^{206}Pb^*$	±	$^{207}Pb^*$	±	$^{206}Pb^*$	±
(ppm)	^{204}Pb	^{235}U	(%)	^{238}U	(%)	corr.	^{238}U	(Ma)	^{235}U	(Ma)	$^{207}Pb^*$	(Ma)
Lee Formation, Little Stone Gap of western Virginia (cont.)												
15	40692	12.8717	1.76	0.4800	1.61	0.9	2527.3	49.6	2670.3	206.9	*2780.5*	*5.8*
84	27396	12.5822	0.91	0.4839	0.90	1.0	2544.3	27.9	2648.9	110.6	*2729.8*	*1.5*
Raccoon Mountain Formation, Rock Mountain syncline, northwestern Georgia												
20	9465	0.5451	10.09	0.0660	1.27	0.1	*411.9*	*5.4*	441.8	54.4	601.0	108.4
42	4416	0.5659	4.73	0.0722	1.33	0.3	*449.7*	*6.2*	455.4	26.8	484.2	50.1
14	7719	0.5981	9.35	0.0726	1.45	0.2	*451.7*	*6.8*	476.1	55.2	595.2	100.0
67	17409	0.6170	2.53	0.0744	0.98	0.4	*462.8*	*4.7*	487.9	15.7	607.7	25.2
49	3977	0.6219	4.48	0.0761	0.66	0.1	*472.9*	*3.2*	491.1	27.9	576.9	48.2
152	35100	0.6851	1.76	0.0804	1.43	0.8	*498.6*	*7.4*	529.9	12.2	666.9	11.0
60	126538	0.8303	1.89	0.0995	0.68	0.4	*611.7*	*4.4*	613.8	15.8	621.6	19.1
59	11549	1.7115	1.43	0.1662	1.14	0.8	*991.4*	*12.2*	1012.8	24.5	1059.6	8.7
10	5983	1.8206	4.33	0.1767	1.49	0.3	*1049.0*	*16.9*	1052.9	77.0	1060.9	40.9
20	12237	1.8405	2.71	0.1792	1.14	0.4	*1062.8*	*13.2*	1060.0	49.4	1054.3	24.7
30	9394	1.8751	1.36	0.1803	0.62	0.5	*1068.8*	*7.2*	1072.3	25.6	1079.5	12.2
7	10561	1.9120	6.36	0.1805	1.79	0.3	*1069.7*	*20.8*	1085.3	116.6	1116.8	60.9
26	24317	1.8795	2.17	0.1817	0.76	0.3	*1076.2*	*8.9*	1073.9	40.6	1069.2	20.4
88	119370	1.9196	0.93	0.1818	0.72	0.8	*1076.6*	*8.4*	1087.9	17.9	1110.6	5.9
60	28123	1.9187	1.28	0.1823	1.07	0.8	*1079.4*	*12.5*	1087.6	24.7	1104.0	7.1
38	25774	1.9597	1.64	0.1826	0.93	0.6	*1081.2*	*10.9*	1101.8	32.1	1142.7	13.4
30	76947	1.8888	2.18	0.1835	1.07	0.5	*1085.9*	*12.6*	1077.2	41.0	1059.5	19.1
15	11798	1.9834	2.89	0.1839	0.86	0.3	*1088.0*	*10.2*	1109.9	56.6	1153.0	27.4
44	429378	1.8961	1.40	0.1842	0.75	0.5	*1089.9*	*8.9*	1079.7	26.6	1059.3	11.9
21	4941	2.0010	2.95	0.1851	1.55	0.5	*1094.8*	*18.4*	1115.9	58.3	1157.2	24.9
28	115026	1.9353	1.66	0.1866	0.52	0.3	*1102.9*	*6.2*	1093.4	32.1	1074.4	15.8
57	20901	1.9737	1.79	0.1874	1.37	0.8	*1107.1*	*16.5*	1106.6	35.3	1105.5	11.6
42	31642	2.0513	1.61	0.1890	1.35	0.8	*1115.9*	*16.4*	1132.7	33.0	1165.2	8.7
16	17690	1.9872	3.50	0.1897	1.42	0.4	*1119.8*	*17.4*	1111.2	68.4	1094.3	32.1
31	7350	2.0878	1.68	0.1914	0.81	0.5	*1128.7*	*10.0*	1144.8	35.0	1175.4	14.6
109	10097	2.2693	1.37	0.1927	1.03	0.8	*1135.9*	*12.7*	1202.8	31.0	1325.1	8.7
42	24529	2.2130	1.08	0.1952	0.59	0.5	*1149.3*	*7.5*	1185.1	24.1	1251.1	8.9
13	11598	2.0995	5.04	0.1955	1.54	0.3	*1151.3*	*19.4*	1148.6	102.1	1143.7	47.7
194	8133	2.1794	1.15	0.1957	1.08	0.9	*1152.2*	*13.6*	1174.5	25.1	1215.8	3.8
87	412654	2.1090	1.12	0.1958	1.02	0.9	*1152.7*	*12.9*	1151.7	23.7	1149.9	4.5
8	2417	2.0671	6.18	0.1963	0.74	0.1	*1155.3*	*9.4*	1138.0	122.1	1105.1	61.3
18	11817	2.1724	2.46	0.1965	1.27	0.5	*1156.4*	*16.1*	1172.2	52.8	1201.6	20.7
11	15613	2.2817	4.68	0.1986	0.77	0.2	*1167.8*	*9.9*	1206.6	103.0	1276.9	45.0
39	25976	2.1859	1.39	0.2000	0.73	0.5	*1175.4*	*9.5*	1176.6	30.4	1178.7	11.7
33	22494	2.2054	1.83	0.2001	1.19	0.7	*1175.8*	*15.4*	1182.8	40.1	1195.6	13.7
27	48441	2.2697	2.33	0.2008	0.97	0.4	*1179.8*	*12.5*	1202.9	52.4	1244.7	20.8
16	17264	2.2455	2.37	0.2010	1.40	0.6	*1180.6*	*18.1*	1195.4	52.7	1222.1	18.8
72	46826	2.2396	1.04	0.2039	0.86	0.8	*1196.2*	*11.2*	1193.5	23.5	1188.7	5.9
119	34442	2.3830	1.17	0.2040	1.11	0.9	*1196.8*	*14.5*	1237.5	28.0	1309.2	3.8
38	13337	2.2565	1.50	0.2050	1.06	0.7	*1202.2*	*14.0*	1198.8	33.9	1192.7	10.5
115	821501	2.3701	0.53	0.2067	0.42	0.8	1211.4	5.6	1233.6	12.8	*1272.6*	*3.2*

Appendix A (cont.). U-Pb geochronologic analyses by Laser-Ablation Multicollector ICP Mass Spectrometry

		Isotopic ratios							Apparent ages (Ma)				
U (ppm)	$^{206}Pb/^{204}Pb$	$^{207}Pb^*/^{235}U$	± (%)	$^{206}Pb^*/^{238}U$	± (%)	error corr.	$^{206}Pb^*/^{238}U$	± (Ma)	$^{207}Pb^*/^{235}U$	± (Ma)	$^{206}Pb^*/^{207}Pb^*$	± (Ma)
Raccoon Mountain Formation, Rock Mountain syncline, northwestern Georgia (cont.)												
27	10376	2.6571	2.35	0.2071	1.11	0.5	1213.2	14.8	1316.6	61.4	*1489.3*	*19.6*
20	20168	2.4277	2.18	0.2073	0.50	0.2	1214.4	6.7	1250.9	52.3	*1314.2*	*20.6*
29	11345	2.2760	1.30	0.2073	0.62	0.5	*1214.6*	*8.3*	1204.9	29.5	1187.5	11.2
48	45338	2.3865	1.55	0.2074	1.48	1.0	1215.1	19.7	1238.5	36.9	*1279.6*	*4.7*
44	20512	2.2883	1.08	0.2079	0.36	0.3	*1217.7*	*4.9*	1208.7	24.8	1192.7	10.1
92	20100	2.3870	1.22	0.2091	1.17	1.0	1224.0	15.8	1238.7	29.1	*1264.4*	*3.2*
26	5938	2.4223	1.93	0.2137	1.25	0.6	1248.3	17.1	1249.2	46.4	*1250.8*	*14.4*
37	16771	2.3951	1.70	0.2137	1.35	0.8	1248.7	18.5	1241.1	40.5	*1228.0*	*10.2*
11	60377	2.7066	4.02	0.2268	0.74	0.2	1317.7	10.8	1330.3	104.9	*1350.5*	*38.1*
19	24406	2.8703	2.28	0.2365	1.43	0.6	1368.7	21.8	1374.1	64.3	*1382.6*	*17.0*
27	47063	3.0576	1.26	0.2438	0.88	0.7	1406.5	13.9	1422.1	38.5	*1445.6*	*8.6*
48	1022	3.2110	1.08	0.2504	0.78	0.7	1440.5	12.6	1459.8	34.7	*1488.1*	*7.1*
20	6944	3.6992	1.47	0.2556	1.14	0.8	1467.1	18.8	1571.2	53.7	*1714.0*	*8.5*
56	22102	4.2010	1.15	0.2844	1.08	0.9	1613.7	19.8	1674.2	48.1	*1750.9*	*3.6*
8	49658	4.1778	3.42	0.2941	1.33	0.4	1661.9	25.2	1669.7	135.7	*1679.5*	*29.1*
10	12922	4.2115	2.95	0.2969	1.38	0.5	1676.0	26.3	1676.3	119.0	*1676.6*	*24.1*
17	22926	4.5701	1.32	0.3146	0.56	0.4	1763.3	11.4	1743.8	59.5	*1720.6*	*11.0*
60	43649	6.7524	1.16	0.3221	1.02	0.9	1799.8	21.2	2079.5	76.9	*2369.2*	*4.8*
18	20470	5.1044	1.57	0.3313	1.21	0.8	1844.6	25.8	1836.8	78.2	*1828.0*	*9.0*
39	43652	5.1785	1.07	0.3332	0.95	0.9	1853.7	20.3	1849.1	55.0	*1843.9*	*4.6*
31	31404	5.2475	1.24	0.3369	1.10	0.9	1871.9	23.8	1860.4	63.8	*1847.5*	*5.1*
41	15557	5.1954	0.75	0.3373	0.55	0.7	1873.7	12.0	1851.9	38.6	*1827.5*	*4.5*
48	20368	9.5915	1.85	0.4091	1.82	1.0	2210.8	47.9	2396.4	166.0	*2558.1*	*2.6*
8	20027	13.1217	1.92	0.4797	1.65	0.9	2526.0	50.8	2688.4	228.3	*2812.9*	*8.1*
9	37657	13.7022	2.27	0.5165	2.20	1.0	2684.2	73.0	2729.3	274.9	*2762.9*	*4.4*
29	10119	14.8685	1.02	0.5451	0.99	1.0	2804.9	34.6	2806.9	143.6	*2808.3*	*2.2*
75	12337	14.9947	0.97	0.5474	0.79	0.8	2814.5	27.7	2814.9	138.2	*2815.2*	*4.7*
Montevallo coal zone conglomerate, Pottsville Formation, central Alabama												
17	2391.981	1.433563	3.30785	0.15536	0.826	0.25	*931.0*	8.3	903.0	47.0	835.3	33.4
120	2822.63	1.686883	1.00719	0.15766	0.837	0.831	*943.8*	8.5	1003.6	17.1	1136.6	5.6
29	5879.304	1.745981	1.30427	0.1702	0.508	0.389	*1013.2*	5.6	1025.7	22.9	1052.4	12.1
90	17615.39	1.772801	0.98176	0.17071	0.545	0.556	*1016.0*	6.0	1035.5	17.5	1076.9	8.2
29	24847.12	1.754704	1.92053	0.17095	0.711	0.37	*1017.4*	7.8	1028.9	33.7	1053.5	18.0
13	1641.154	1.598549	4.193	0.17368	0.98	0.234	*1032.4*	11.0	969.6	65.9	830.1	42.5
16	17083.96	1.759337	3.10777	0.17406	1.169	0.376	*1034.4*	13.1	1030.6	54.1	1022.5	29.1
15	6892.95	1.779888	2.80826	0.17548	0.746	0.266	*1042.2*	8.4	1038.1	49.5	*1029.5*	27.4
13	3191.722	1.838594	4.14499	0.17646	0.475	0.115	*1047.6*	5.4	1059.4	74.6	1083.7	41.3
20	5960.463	1.802215	2.144	0.17829	0.618	0.288	*1057.6*	7.1	1046.3	38.5	1022.6	20.8
15	1906.866	1.912402	2.46391	0.17941	0.737	0.299	*1063.7*	8.5	1085.4	46.8	1129.2	23.4
12	2999.85	1.676994	3.3414	0.18213	0.634	0.19	*1078.6*	7.4	999.8	55.4	830.9	34.2
26	11466.07	1.840669	1.75078	0.1822	1.034	0.591	*1079.0*	12.1	1060.1	32.2	1021.5	14.3
62	2417.178	1.956678	1.59088	0.18279	0.773	0.486	*1082.2*	9.1	1100.7	31.1	1137.6	13.8
9	2061.716	1.905953	4.10344	0.18813	1.404	0.342	*1111.3*	17.0	1083.2	76.5	1027.1	39.0
29	7062.32	2.081099	2.21488	0.18822	1.116	0.504	*1111.7*	13.5	1142.6	45.8	1201.7	18.9

Appendix A (cont.). U-Pb geochronologic analyses by Laser-Ablation Multicollector ICP Mass Spectrometry

	Isotopic ratios						Apparent ages (Ma)					
U (ppm)	^{206}Pb / ^{204}Pb	^{207}Pb* / ^{235}U	± (%)	^{206}Pb* / ^{238}U	± (%)	error corr.	^{206}Pb* / ^{238}U	± (Ma)	^{207}Pb* / ^{235}U	± (Ma)	^{206}Pb* / ^{207}Pb*	± (Ma)
Montevallo coal zone conglomerate, Pottsville Formation, central Alabama (cont.)												
110	18411.35	1.988849	0.85606	0.18852	0.719	0.84	*1113.4*	8.7	1111.7	17.1	1108.5	4.6
9	9794.772	1.988815	3.97515	0.18917	0.948	0.238	*1116.9*	11.5	1111.7	77.3	1101.7	38.6
16	33017.26	2.079934	3.04415	0.19005	0.394	0.13	*1121.7*	4.8	1142.2	62.3	1181.5	29.8
14	19579.9	2.164396	4.70558	0.19285	1.149	0.244	*1136.8*	14.3	1169.7	98.5	1231.1	44.8
62	21710.95	2.205877	0.97736	0.19891	0.837	0.857	*1169.5*	10.7	1182.9	21.7	1207.6	5.0
14	4796.606	2.13943	1.92918	0.19994	1.041	0.54	*1175.0*	13.4	1161.6	41.1	1136.8	16.2
61	18998.7	2.313777	0.94709	0.2023	0.749	0.79	*1187.7*	9.8	1216.5	22.0	1268.1	5.7
13	6088.457	2.224231	2.90621	0.20233	1.835	0.631	*1187.8*	23.9	1188.7	63.6	1190.3	22.2
22	4917.184	2.310752	2.11506	0.20998	0.966	0.457	*1228.7*	13.1	1215.6	48.5	1192.3	18.6
105	4339.783	2.540711	1.40504	0.2131	1.334	0.949	1245.3	18.3	1283.8	35.6	*1348.7*	*4.3*
14	6415.445	2.644951	2.75154	0.22487	1.194	0.434	1307.6	17.3	1313.2	71.3	*1322.5*	*24.0*
11	3090.665	2.935743	3.3069	0.22578	0.959	0.29	1312.4	13.9	1391.2	94.1	*1514.2*	*29.9*
15	9202.068	2.888438	2.53901	0.23598	0.439	0.173	1365.8	6.7	1378.9	71.9	*1399.3*	*24.0*
37	16448.73	2.856485	1.06122	0.23721	0.605	0.57	1372.2	9.2	1370.5	30.3	*1367.9*	*8.4*
40	16099.65	3.081331	0.98876	0.23815	0.719	0.727	1377.1	11.0	1428.1	30.5	*1504.9*	*6.4*
32	6471.163	3.069711	1.01129	0.24771	0.556	0.55	1426.6	8.9	1425.2	31.0	*1422.9*	*8.1*
14	7581.864	3.548289	1.95307	0.25709	0.765	0.392	1474.9	12.7	1538.1	68.0	*1626.0*	*16.7*
5	2434.681	3.736458	4.86351	0.26807	1.352	0.278	1531.0	23.3	1579.2	169.5	*1644.2*	*43.3*
42	2796.932	4.256564	1.2661	0.27796	1.051	0.83	1581.1	18.8	1685.0	53.3	*1816.9*	*6.4*
7	20696.23	15.63652	1.0386	0.54866	0.711	0.684	2819.6	25.1	2854.9	152.8	*2879.8*	*6.2*

U concentration has an uncertainty of ~25%.
Decay constants: ^{235}U=9.8485x10-10. ^{238}U=1.55125x10-10, ^{238}U/^{235}U=137.88.
Uncertainties include only random errors, and are shown at the 1-sigma level.
Isotope ratios are corrected for Pb/U fractionation by comparison with
standard zircon with an age of 564 ± 4 Ma.
Initial Pb composition interpreted from Stacey and Kramers (1975), with uncertainties of
1.0 for ^{206}Pb/^{204}Pb and 0.3 for ^{207}Pb/^{204}Pb.
*Italicized numbers refer to best ages.

Appendix B. LA-ICPMS U-Pb isotopic data for zircons from two Permian sandstones, Appalachian basin.

Analysis[1] #	Color/ morphology[2]	U (ppm)	Th (ppm)	U/Th	^{207}Pb/^{235}U	±(%)	208Pb/ 206Pb	±(%)	errcorr	207Pb/ 206Pb	±(%)	^{206}Pb/ ^{238}U age (Ma)	±(Ma)	^{206}Pb/ ^{207}Pb age (Ma)	±(Ma)	Apparent ages[3] (Ma)	±(Ma)
Greene Formation, Holbrook 7.5-minute quadrangle, UTM N17 (N563406, E563406) Pennsylvania																	
67	opaque, facet	1780	441	4.0	0.512	13.8	0.0485	5.2	0.377	13.07	12.8	305.4	15.5	1108.4	127.8	305.4	15.5
102	pink, facet	227	209	1.1	0.616	22.0	0.0829	5.3	0.242	18.54	21.3	513.3	26.2	368.1	240.2	513.3	26.2
93	yllw, round	1031	215	4.8	1.193	6.7	0.1132	3.4	0.503	13.09	5.8	691.4	21.9	1105.4	57.5	691.4	21.9
44	pink, round	43	89	0.5	1.449	38.3	0.1364	8.0	0.210	12.98	37.4	824.5	61.9	1122.0	373.2	824.5	61.9
99	pink, facet	30	17	1.8	1.659	46.7	0.1419	8.3	0.179	11.80	46.0	855.6	66.5	1310.0	445.9	855.6	66.5
9	pink, round	73	74	1.0	1.369	33.6	0.1511	2.2	0.067	15.22	33.5	906.9	18.9	797.5	351.2	906.9	18.9
7	pink, round	603	867	0.7	1.570	11.2	0.1542	2.3	0.202	13.54	11.0	924.4	19.4	1037.2	110.6	924.4	19.4
100	pink, facet	1012	188	5.4	1.607	10.1	0.1544	9.0	0.892	13.25	4.6	925.4	77.0	1081.3	45.7	925.4	77.0
13	pink, round	164	120	1.4	1.959	11.5	0.1636	5.3	0.458	11.51	10.3	976.6	47.7	1357.3	98.9	976.6	47.7
29	pink, round	50	31	1.6	1.984	15.9	0.1719	3.4	0.217	11.95	15.5	1022.7	32.4	1285.4	150.8	1022.7	32.4
23	pink, round	142	55	2.6	1.713	18.4	0.1729	3.1	0.167	13.92	18.2	1028.3	29.2	981.5	185.2	1028.3	29.2
5	pink, round	486	99	4.9	1.643	10.3	0.1743	4.5	0.430	14.63	9.3	1035.6	42.4	879.4	96.6	1035.6	42.4
72	opaque, round	818	310	2.6	1.766	9.6	0.1744	2.2	0.234	13.61	9.3	1036.3	21.4	1026.4	94.4	1036.3	21.4
49	pink, round	101	35	2.8	1.923	12.8	0.1760	3.1	0.239	12.62	12.5	1045.0	29.5	1178.1	123.2	1045.0	29.5
66	opaque, round	387	52	7.4	1.814	7.6	0.1789	3.9	0.515	13.59	6.5	1060.9	38.1	1029.5	65.7	1060.9	38.1
40	pink, round	345	77	4.5	1.718	6.1	0.1797	3.1	0.510	14.42	5.3	1065.3	30.6	908.7	54.2	1065.3	30.6
15	pink, round	573	140	4.1	1.605	13.9	0.1801	3.5	0.249	15.47	13.5	1067.4	34.1	762.9	142.2	1067.4	34.1
19	pink, round	543	75	7.2	1.791	8.7	0.1809	6.0	0.696	13.92	6.2	1071.6	59.4	980.8	63.4	1071.6	59.4
55	pink, round	109	60	1.8	1.824	18.2	0.1816	3.8	0.211	13.73	17.8	1075.6	38.0	1009.9	180.4	1075.6	38.0
21	pink, round	72	36	2.0	1.696	25.9	0.1817	3.4	0.130	14.78	25.7	1076.5	33.4	858.4	266.5	1076.5	33.4
110	pink, round	82	26	3.1	1.847	20.6	0.1818	5.5	0.269	13.57	19.9	1076.6	54.7	1033.4	200.8	1076.6	54.7
89	yllw, round	174	61	2.9	1.819	8.9	0.1818	2.6	0.289	13.78	8.5	1076.7	25.5	1001.8	86.6	1076.7	25.5
6	pink, round	50	19	2.7	1.476	37.2	0.1831	3.6	0.096	17.11	37.0	1084.0	35.5	546.9	404.5	1084.0	35.5
85	yllw, round	56	28	2.0	1.316	43.6	0.1832	2.7	0.063	19.19	43.5	1084.5	27.3	290.0	497.0	1084.5	27.3
88	yllw, round	38	16	2.3	1.808	36.9	0.1833	4.3	0.116	13.98	36.6	1085.0	42.5	972.7	373.5	1085.0	42.5
17	pink, round	221	94	2.3	1.996	7.0	0.1849	2.4	0.335	12.77	6.6	1093.7	23.6	1154.6	65.7	1093.7	23.6
14	pink, round	341	118	2.9	1.795	7.5	0.1854	4.2	0.554	14.24	6.2	1096.2	41.7	935.2	63.9	1096.2	41.7
92	yllw, round	94	23	4.2	2.140	11.8	0.1854	5.5	0.462	11.94	10.5	1096.2	54.8	1286.1	102.0	1096.2	54.8
52	pink, round	422	255	1.7	2.117	5.9	0.1864	1.2	0.207	12.14	5.8	1102.0	12.4	1254.3	56.3	1102.0	12.4

Appendix B (cont.). LA-ICPMS U-Pb isotopic data for zircons from two Permian sandstones, Appalachian basin.

Analysis[1] #	Color/ morphology[2]	U (ppm)	Th (ppm)	U/Th	^{207}Pb/^{235}U	±(%)	208Pb/ 206Pb	±(%)	errcorr	207Pb/ 206Pb	±(%)	^{206}Pb/ ^{238}U age (Ma)	±(Ma)	^{206}Pb/ ^{207}Pb age (Ma)	±(Ma)	Apparent ages[3] (Ma)	±(Ma)
Greene Formation, Holbrook 7.5-minute quadrangle, UTM N17 (N563406, E563406) Pennsylvania																	
28	pink, round	161	188	0.9	2.019	7.8	0.1870	2.7	0.344	12.77	7.3	1105.1	27.1	1155.0	72.4	1105.1	27.1
103	pink, round	175	41	4.3	2.043	5.7	0.1878	3.2	0.561	12.68	4.7	1109.4	32.6	1169.5	46.9	1109.4	32.6
56	pink, round	58	52	1.1	1.572	30.0	0.1879	5.3	0.176	16.48	29.5	1110.0	53.6	628.1	318.1	1110.0	53.6
97	yllw, round	72	59	1.2	1.858	12.7	0.1893	4.3	0.337	14.04	12.0	1117.5	43.9	963.2	122.6	1117.5	43.9
53	pink, round	137	67	2.0	1.960	11.5	0.1897	4.5	0.389	13.35	10.6	1120.0	46.0	1065.9	106.8	1120.0	46.0
27	pink, round	310	137	2.3	1.990	6.3	0.1910	4.6	0.735	13.24	4.3	1126.8	47.8	1083.1	42.9	1126.8	47.8
16	pink, round	287	211	1.4	2.018	14.4	0.1923	6.0	0.414	13.14	13.2	1134.0	61.9	1098.0	131.6	1134.0	61.9
12	pink, round	177	70	2.5	2.041	8.7	0.1926	2.0	0.229	13.01	8.5	1135.4	20.7	1117.5	84.6	1135.4	20.7
39	pink, round	182	46	4.0	1.760	15.9	0.1933	5.6	0.355	15.14	14.9	1139.3	58.7	807.8	155.4	1139.3	58.7
104	pink, round	56	52	1.1	1.863	19.2	0.1942	4.1	0.212	14.37	18.7	1143.9	42.5	916.7	192.6	1143.9	42.5
18	pink, round	96	43	2.2	1.858	19.0	0.1952	7.4	0.391	14.49	17.5	1149.6	77.9	899.4	180.6	1149.6	77.9
32	pink, round	67	21	3.2	1.932	27.4	0.1972	2.6	0.093	14.07	27.3	1160.4	27.1	959.1	279.1	1160.4	27.1
98	pink, facet	185	100	1.9	2.103	7.3	0.1986	3.3	0.454	13.02	6.5	1167.9	35.3	1115.5	64.8	1167.9	35.3
45	pink, round	334	194	1.7	2.082	5.0	0.1988	2.2	0.434	13.16	4.5	1169.0	22.9	1094.0	44.7	1169.0	22.9
1	pink, round	170	120	1.4	2.304	7.3	0.2118	4.3	0.595	12.68	5.8	1238.5	48.5	1169.1	57.7	1169.1	57.7
33	pink, round	252	71	3.6	2.214	6.4	0.2032	2.5	0.392	12.66	5.9	1192.5	27.1	1172.4	57.9	1192.5	27.1
101	pink, facet	210	22	9.6	1.977	14.7	0.2049	6.1	0.417	14.30	13.4	1201.8	66.9	927.0	137.2	1201.8	66.9
84	yllw, round	418	382	1.1	2.277	9.9	0.2066	3.1	0.317	12.51	9.4	1210.6	34.5	1195.6	92.6	1210.6	34.5
74	opaque, round	376	117	3.2	2.300	14.4	0.2067	4.1	0.284	12.39	13.8	1211.4	45.0	1213.7	136.1	1211.4	45.0
42	pink, round	132	46	2.9	2.261	8.9	0.2072	4.0	0.445	12.63	8.0	1213.7	43.8	1176.2	79.0	1213.7	43.8
83	yllw, round	144	76	1.9	1.915	21.4	0.2085	4.7	0.219	15.01	20.9	1220.7	51.9	825.6	217.6	1220.7	51.9
11	pink, round	206	31	6.7	2.177	9.0	0.2091	5.3	0.592	13.24	7.2	1223.8	58.9	1082.7	72.6	1223.8	58.9
58	pink, round	89	37	2.4	2.223	17.5	0.2103	2.5	0.143	13.04	17.3	1230.4	28.0	1112.7	172.4	1230.4	28.0
25	pink, round	385	202	1.9	2.282	4.5	0.2106	3.3	0.735	12.73	3.1	1232.1	37.1	1161.1	30.4	1232.1	37.1
41	pink, round	85	50	1.7	2.321	13.9	0.2148	1.9	0.137	12.76	13.8	1254.3	21.6	1156.6	136.6	1254.3	21.6
96	yllw, round	348	133	2.6	2.549	5.1	0.2228	4.0	0.780	12.05	3.2	1296.7	46.9	1268.8	31.4	1268.8	31.4
57	pink, round	102	48	2.1	2.692	8.9	0.2346	2.9	0.326	12.02	8.5	1358.6	35.6	1274.4	82.4	1274.4	82.4
30	pink, round	595	312	1.9	2.544	4.1	0.2214	3.6	0.888	12.00	1.9	1289.1	42.0	1277.2	18.2	1277.2	18.2
107	pink, round	107	76	1.4	2.667	10.2	0.2316	3.2	0.317	11.97	9.7	1342.8	39.2	1281.9	94.5	1281.9	94.5
47	pink, round	97	160	0.6	2.357	9.1	0.2040	2.2	0.247	11.94	8.8	1196.9	24.5	1287.3	85.7	1287.3	85.7
20	pink, round	471	414	1.1	2.333	11.8	0.2016	9.3	0.791	11.91	7.2	1183.9	100.1	1291.4	70.2	1291.4	70.2

Appendix B (cont.). LA-ICPMS U-Pb isotopic data for zircons from two Permian sandstones, Appalachian basin.

Greene Formation, Holbrook 7.5-minute quadrangle, UTM N17 (N563406, E563406) Pennsylvania

Analysis[1] #	Color/ morphology[2]	U (ppm)	Th (ppm)	U/Th	$^{207}Pb/^{235}U$	±(%)	208Pb/ 206Pb	±(%)	errcorr	207Pb/ 206Pb	±(%)	$^{206}Pb/^{238}U$ age (Ma)	±(Ma)	$^{206}Pb/^{207}Pb$ age (Ma)	±(Ma)	Apparent ages[3] (Ma)	Apparent ages[3] ±(Ma)
94	yllw, round	108	86	1.3	2.629	14.5	0.2223	4.6	0.319	11.66	13.7	1294.2	53.9	1333.0	132.7	1294.2	53.9
105	pink, round	78	74	1.1	2.238	19.6	0.2243	5.0	0.253	13.82	19.0	1304.6	58.4	996.3	193.1	1304.6	58.4
91	yllw, round	139	153	0.9	2.489	8.6	0.2122	2.7	0.318	11.76	8.2	1240.6	30.9	1316.8	79.3	1316.8	79.3
10	pink, round	73	35	2.1	2.284	12.5	0.1941	5.5	0.439	11.72	11.3	1143.6	57.3	1323.6	109.0	1323.6	109.0
26	pink, round	99	92	1.1	2.555	7.9	0.2150	2.7	0.337	11.60	7.5	1255.5	30.4	1342.3	72.2	1342.3	72.2
48	pink, round	74	22	3.4	2.108	16.4	0.1754	3.0	0.186	11.47	16.1	1041.9	29.2	1363.7	154.9	1363.7	154.9
87	yllw, round	159	72	2.2	3.009	9.9	0.2496	4.3	0.437	11.44	8.9	1436.3	55.7	1370.3	86.0	1370.3	86.0
90	yllw, round	67	23	3.0	2.445	19.5	0.1986	3.7	0.188	11.20	19.1	1168.0	38.9	1410.3	183.0	1410.3	183.0
2	pink, round	50	17	3.0	2.678	10.9	0.2175	2.6	0.235	11.20	10.6	1268.4	29.4	1410.9	101.4	1410.9	101.4
51	pink, round	96	54	1.8	2.596	20.8	0.2458	2.8	0.135	13.05	20.6	1416.8	35.7	1110.9	206.0	1416.8	35.7
109	pink, round	114	85	1.3	3.211	5.9	0.2592	2.5	0.421	11.13	5.4	1485.9	33.0	1421.9	51.3	1421.9	51.3
34	pink, round	248	111	2.2	3.081	4.5	0.2481	2.7	0.604	11.10	3.6	1428.5	34.5	1427.5	34.0	1427.5	34.0
24	pink, round	82	40	2.1	3.395	7.4	0.2718	2.5	0.340	11.04	7.0	1549.8	34.6	1438.4	66.4	1438.4	66.4
43	pink, round	180	91	2.0	2.756	5.6	0.2204	1.7	0.300	11.03	5.3	1284.0	19.4	1439.9	50.6	1439.9	50.6
22	pink, round	407	58	7.0	3.303	3.1	0.2607	1.8	0.574	10.89	2.6	1493.6	24.0	1464.6	24.4	1464.6	24.4
46	pink, round	117	69	1.7	3.918	8.5	0.3029	3.0	0.360	10.66	7.9	1705.6	45.4	1504.5	74.5	1504.5	74.5
106	pink, round	42	40	1.0	2.790	18.4	0.2153	4.7	0.255	10.64	17.8	1256.9	53.3	1507.8	167.8	1507.8	167.8
95	yllw, round	161	102	1.6	3.603	6.5	0.2749	4.9	0.750	10.52	4.3	1565.6	67.5	1529.2	40.5	1529.2	40.5
8	pink, round	194	129	1.5	3.452	7.6	0.2508	5.2	0.681	10.02	5.5	1442.7	66.2	1620.7	51.5	1620.7	51.5
108	pink, round	66	61	1.1	8.652	7.1	0.4503	3.5	0.502	7.18	6.1	2396.5	70.5	2219.3	52.9	2219.3	52.9

Washington Formation, Parkersburg 7.5-minute quadrangle UTM N17(N434869, E453707), Ohio

Analysis[1] #	Color/ morphology[2]	U (ppm)	Th (ppm)	U/Th	$^{207}Pb/^{235}U$	±(%)	208Pb/ 206Pb	±(%)	errcorr	207Pb/ 206Pb	±(%)	$^{206}Pb/^{238}U$ age (Ma)	±(Ma)	$^{206}Pb/^{207}Pb$ age (Ma)	±(Ma)	Apparent ages[3] (Ma)	Apparent ages[3] ±(Ma)
23	pink, ac facet	752	779	1.0	0.4	26.2	0.0	4.3	0.2	17.4	25.8	314.5	13.1	514.2	283.7	314.5	13.1
6	pink, ac facet	353	160	2.2	0.4	16.6	0.1	3.4	0.2	17.2	16.3	348.5	11.6	531.5	178.3	348.5	11.6
16	pink, ac facet	383	296	1.3	0.4	25.2	0.1	3.0	0.1	20.4	25.0	354.9	10.3	143.7	293.9	354.9	10.3
105	yl-pk, round	493	102	4.8	0.5	18.4	0.1	8.0	0.4	17.2	16.5	364.8	28.4	538.0	180.8	364.8	28.4

Appendix B (cont.). LA-ICPMS U-Pb isotopic data for zircons from two Permian sandstones, Appalachian basin.

Analysis[1] #	Color/ morphology[2]	U (ppm)	Th (ppm)	U/Th	^{207}Pb/^{235}U	±(%)	208Pb/ 206Pb	±(%)	errcorr	207Pb/ 206Pb	±(%)	^{206}Pb/ ^{238}U age (Ma)	±(Ma)	^{206}Pb/ ^{207}Pb age (Ma)	±(Ma)	Apparent ages[3] (Ma)	±(Ma)
Washington Formation, Parkersburg 7.5-minute quadrangle UTM N17(N434869, E453707), Ohio																	
2	pink, ac facet	109	75	1.4	0.4	50.5	0.1	4.6	0.1	19.9	50.3	373.6	16.7	210.6	582.4	373.6	16.7
18	pink, ac facet	279	415	0.7	0.4	22.0	0.1	1.9	0.1	19.3	21.9	384.4	7.1	278.9	250.8	384.4	7.1
95	yl-pk, round	241	278	0.9	0.5	22.0	0.1	5.5	0.3	18.4	21.3	388.7	20.8	382.9	239.4	388.7	20.8
8	pink, ac facet	852	258	3.3	0.5	9.1	0.1	5.7	0.6	18.3	7.2	393.2	21.5	403.8	80.3	393.2	21.5
21	pink, ac facet	188	200	0.9	0.5	18.0	0.1	4.1	0.2	17.0	17.5	396.0	15.8	554.5	190.9	396.0	15.8
11	pink, ac facet	133	147	0.9	0.5	31.6	0.1	4.1	0.1	18.5	31.3	421.7	16.8	374.6	352.1	421.7	16.8
26	pink, round	93	141	0.7	0.9	29.1	0.1	3.9	0.1	10.0	28.8	424.8	16.2	1632.8	268.0	424.8	16.2
22	pink, ac facet	292	60	4.9	0.6	15.6	0.1	6.6	0.4	17.3	14.2	432.2	27.4	526.8	155.6	432.2	27.4
12	pink, ac facet	80	126	0.6	0.7	28.7	0.1	3.4	0.1	14.3	28.5	444.6	14.8	923.5	292.8	444.6	14.8
35r	opaque, round	1319	2243	0.6	1.0	18.5	0.1	11.7	0.6	12.6	14.3	541.4	60.3	1182.5	141.5	541.4	60.3
97	yl-pk, round	253	169	1.5	0.8	13.6	0.1	3.7	0.3	16.1	13.1	564.0	20.0	676.5	140.4	564.0	20.0
17	pink, ac facet	217	259	0.8	0.9	9.0	0.1	2.2	0.2	16.3	8.8	649.9	13.3	653.6	94.0	649.9	13.3
94	yl-pk, round	50	66	0.8	1.3	24.1	0.1	9.2	0.4	11.6	22.3	676.7	58.8	1336.4	215.6	676.7	58.8
48	opaque, round	2004	1751	1.1	1.3	10.2	0.1	2.7	0.3	12.9	9.8	713.9	18.3	1140.7	97.7	713.9	18.3
30	opaque, round	1267	1276	1.0	1.2	13.1	0.1	10.4	0.8	13.9	8.0	729.2	71.5	988.6	80.9	729.2	71.5
88r	yl-pk, round	80	142	0.6	1.4	7.0	0.1	1.7	0.2	12.3	6.7	745.2	11.9	1232.0	66.2	745.2	11.9
42r	opaque, round	687	232	3.0	1.2	3.5	0.1	1.5	0.4	14.2	3.1	778.0	11.3	938.6	32.0	778.0	11.3
99	yl-pk, round	83	35	2.3	1.7	18.6	0.1	4.6	0.2	12.3	18.0	895.5	38.3	1233.6	176.7	895.5	38.3
100	yl-pk, round	370	164	2.3	1.5	5.0	0.2	2.4	0.5	14.0	4.4	909.0	20.5	975.3	45.0	909.0	20.5
5	pink, ac facet	50	32	1.6	1.5	21.8	0.2	4.8	0.2	14.4	21.3	921.4	41.4	912.9	219.1	921.4	41.4
72	pink, round	533	200	2.7	1.6	10.1	0.2	1.5	0.1	13.5	10.0	931.7	13.0	1048.4	100.3	931.7	13.0
67	pink, round	527	1252	0.4	1.6	10.1	0.2	1.7	0.2	13.4	10.0	945.2	14.7	1057.0	100.3	945.2	14.7
20	pink, ac facet	34	25	1.4	1.9	27.8	0.2	2.5	0.1	11.9	27.7	952.3	22.3	1300.6	269.2	952.3	22.3
55	pink, round	877	300	2.9	1.5	8.1	0.2	5.2	0.6	14.5	6.3	953.4	45.7	902.6	64.7	953.4	45.7
96	yl-pk, round	55	24	2.3	0.9	48.0	0.2	6.9	0.1	23.8	47.5	964.6	61.5	-229.6	599.1	964.6	61.5
28	opaque, round	240	112	2.1	1.7	8.2	0.2	4.0	0.5	13.5	7.2	997.3	36.4	1036.4	72.4	997.3	36.4
106	yl-pk, round	194	138	1.4	1.9	9.6	0.2	2.1	0.2	12.4	9.4	1020.9	19.8	1206.6	92.2	1020.9	19.8
51	opaque, round	647	350	1.8	1.8	9.7	0.2	7.2	0.7	13.1	6.5	1029.4	68.1	1102.3	65.1	1029.4	68.1
13	pink, ac facet	58	47	1.2	1.9	18.2	0.2	2.6	0.1	12.4	18.0	1031.3	24.9	1207.8	177.1	1031.3	24.9
102	yl-pk, round	271	110	2.5	1.8	6.7	0.2	5.1	0.8	13.7	4.4	1033.7	48.7	1019.9	44.5	1033.7	48.7

Appendix B (cont.). LA-ICPMS U-Pb isotopic data for zircons from two Permian sandstones, Appalachian basin.

Analysis[1] #	Color/ morphology[2]	U (ppm)	Th (ppm)	U/Th	$^{207}Pb/^{235}U$	±(%)	208Pb/ 206Pb	±(%)	errcorr	207Pb/ 206Pb	±(%)	$^{206}Pb/^{238}U$ age (Ma)	±(Ma)	$^{206}Pb/^{207}Pb$ age (Ma)	±(Ma)	Apparent ages[3] (Ma)	±(Ma)
Washington Formation, Parkersburg 7.5-minute quadrangle UTM N17(N434869, E453707), Ohio																	
93	yl-pk, round	116	57	2.1	2.0	8.8	0.2	3.2	0.4	12.4	8.2	1047.2	31.2	1215.3	80.7	1047.2	31.2
59	pink, round	778	246	3.2	1.7	3.7	0.2	2.2	0.6	14.2	3.0	1048.1	20.9	947.7	30.5	1048.1	20.9
73	pink, round	286	99	2.9	1.8	5.6	0.2	4.4	0.8	13.5	3.4	1049.9	42.6	1038.1	34.0	1049.9	42.6
10	pink, ac facet	91	70	1.3	1.9	8.4	0.2	2.3	0.3	12.6	8.1	1059.1	22.0	1175.6	79.7	1059.1	22.0
75	pink, round	402	248	1.6	1.8	5.8	0.2	3.7	0.6	14.0	4.5	1059.6	35.6	962.9	46.3	1059.6	35.6
89	yl-pk, round	50	40	1.2	1.8	19.6	0.2	6.4	0.3	14.0	18.5	1075.3	63.1	962.8	188.8	1075.3	63.1
14	pink, ac facet	71	45	1.6	2.1	9.1	0.2	2.5	0.3	11.9	8.8	1077.0	24.6	1298.2	85.2	1077.0	24.6
91	yl-pk, round	64	34	1.9	2.1	15.4	0.2	4.8	0.3	12.1	14.6	1080.1	48.0	1266.4	142.5	1080.1	48.0
54	pink, round	235	148	1.6	1.7	16.1	0.2	2.4	0.2	14.7	15.9	1081.7	24.4	863.2	165.3	1081.7	24.4
4	pink, ac facet	96	57	1.7	1.9	13.3	0.2	4.1	0.3	13.6	12.7	1085.8	41.3	1028.3	128.2	1085.8	41.3
85	yl-pk, round	86	50	1.7	1.8	13.6	0.2	3.2	0.2	14.2	13.3	1088.6	32.0	943.1	135.9	1088.6	32.0
24	pink, ac facet	70	34	2.1	2.0	13.2	0.2	4.2	0.3	13.0	12.5	1094.6	42.2	1121.0	124.4	1094.6	42.2
84	yl-pk, round	543	222	2.4	2.0	6.3	0.2	5.8	0.9	12.8	2.5	1096.1	58.0	1148.2	24.4	1096.1	58.0
65	pink, round	459	326	1.4	1.9	5.0	0.2	2.2	0.4	13.8	4.4	1098.9	22.3	993.7	45.2	1098.9	22.3
64	pink, round	143	82	1.7	1.9	8.3	0.2	3.1	0.4	13.5	7.7	1108.8	31.1	1040.1	77.8	1108.8	31.1
3	pink, ac facet	149	109	1.4	1.6	10.7	0.2	2.3	0.2	15.8	10.5	1113.3	23.8	720.1	111.3	1113.3	23.8
1	pink, ac facet	145	154	0.9	1.9	8.2	0.2	1.2	0.1	13.4	8.1	1114.8	11.9	1051.0	81.7	1114.8	11.9
77	pink, round	159	55	2.9	2.3	9.5	0.2	4.2	0.4	13.0	8.6	1290.9	48.6	1117.0	85.4	1117.0	85.4
42c	opaque, round	296	39	7.6	2.0	6.6	0.2	3.1	0.5	13.5	5.8	1135.2	32.2	1041.1	58.0	1135.2	32.2
83	yl-pk, round	46	36	1.3	1.7	26.9	0.2	2.4	0.1	15.5	26.8	1144.3	25.2	764.4	282.7	1144.3	25.2
82	yl-pk, round	99	135	0.7	2.2	15.5	0.2	2.4	0.2	12.3	15.4	1149.8	25.0	1221.8	151.0	1149.8	25.0
88c	yl-pk, round	231	49	4.7	2.1	4.4	0.2	2.2	0.5	13.0	3.8	1160.7	23.4	1117.9	37.7	1160.7	23.4
86	yl-pk, round	439	294	1.5	2.2	6.1	0.2	1.4	0.2	12.6	6.0	1165.4	14.8	1176.4	59.2	1165.4	14.8
60	pink, round	453	194	2.3	2.1	5.7	0.2	5.0	0.9	13.1	2.7	1171.0	53.7	1097.3	27.5	1171.0	53.7
61	pink, round	678	451	1.5	2.2	8.5	0.2	7.4	0.9	12.5	4.2	1176.8	78.9	1194.4	41.4	1176.8	78.9
71	pink, round	278	178	1.6	2.4	7.3	0.2	4.6	0.6	12.5	5.6	1265.2	52.6	1200.6	55.2	1200.6	55.2
45	opaque, round	230	195	1.2	2.2	6.5	0.2	3.8	0.6	12.7	5.3	1204.7	41.8	1158.3	52.6	1204.7	41.8
69	pink, round	179	87	2.1	2.1	6.1	0.2	1.7	0.3	13.3	5.9	1206.1	18.1	1069.8	59.0	1206.1	18.1
15	pink, ac facet	65	29	2.2	2.3	21.9	0.2	4.8	0.2	12.3	21.3	1214.4	52.5	1232.6	209.3	1214.4	52.5
76	pink, round	525	334	1.6	2.3	4.4	0.2	2.5	0.6	12.8	3.6	1230.2	28.3	1148.1	35.4	1230.2	28.3

Appendix B (cont.). LA-ICPMS U-Pb isotopic data for zircons from two Permian sandstones, Appalachian basin.

Analysis[1] #	Color/ morphology[2]	U (ppm)	Th (ppm)	U/Th	$^{207}Pb/^{235}U$	±(%)	208Pb/ 206Pb	±(%)	errcorr	207Pb/ 206Pb	±(%)	$^{206}Pb/^{238}U$ age (Ma)	±(Ma)	$^{206}Pb/^{207}Pb$ age (Ma)	±(Ma)	Apparent ages[3] (Ma)	±(Ma)
Washington Formation, Parkersburg 7.5-minute quadrangle UTM N17(N434869, E453707), Ohio																	
58	pink, round	70	28	2.5	2.3	12.2	0.2	2.4	0.2	12.7	12.0	1235.6	26.5	1163.6	118.7	1235.6	26.5
68	pink, round	83	58	1.4	2.2	11.7	0.2	4.9	0.4	12.1	10.6	1153.2	51.9	1263.2	103.9	1263.2	103.9
87	yl-pk, round	156	62	2.5	2.3	6.4	0.2	3.7	0.6	12.1	5.2	1198.3	40.8	1266.9	50.7	1266.9	50.7
43	opaque, round	267	161	1.7	2.6	6.0	0.2	4.2	0.7	12.0	4.3	1296.5	49.5	1284.7	41.6	1284.7	41.6
107	yl-pk, round	460	299	1.5	2.2	5.6	0.2	4.7	0.8	11.8	3.0	1133.1	48.6	1306.8	29.3	1306.8	29.3
35c	opaque, round	613	480	1.3	2.2	4.0	0.2	2.7	0.7	11.8	3.0	1134.1	28.0	1311.8	28.7	1311.8	28.7
90	yl-pk, round	102	77	1.3	2.5	7.6	0.2	2.1	0.3	11.5	7.3	1228.3	23.5	1353.0	70.4	1353.0	70.4
25	pink, ac facet	79	60	1.3	3.0	9.7	0.2	2.1	0.2	11.5	9.4	1421.7	26.5	1362.2	91.0	1362.2	91.0
66	pink, round	119	70	1.7	3.2	6.9	0.3	3.6	0.5	11.0	5.9	1479.9	47.0	1444.6	56.5	1444.6	56.5
62	pink, round	387	56	6.9	3.3	4.0	0.3	3.8	0.9	10.9	1.3	1504.7	50.9	1455.7	12.8	1455.7	12.8
104	yl-pk, round	82	50	1.7	2.2	14.5	0.2	4.8	0.3	10.9	13.6	1020.2	45.4	1463.3	129.5	1463.3	129.5
81	yl-pk, round	148	166	0.9	3.6	5.3	0.3	3.9	0.7	10.8	3.5	1594.3	55.0	1476.7	33.5	1476.7	33.5
78	yl-pk, round	39	20	2.0	2.0	18.7	0.2	4.5	0.2	10.7	18.1	950.1	39.9	1493.0	171.3	1493.0	171.3
101	yl-pk, round	90	35	2.5	2.6	9.8	0.2	3.1	0.3	10.7	9.3	1186.6	33.1	1493.3	88.4	1493.3	88.4
92	yl-pk, round	95	71	1.3	3.2	10.8	0.2	2.7	0.3	10.4	10.4	1386.8	34.1	1546.1	98.1	1546.1	98.1
50	opaque, round	106	128	0.8	3.3	10.5	0.2	7.4	0.7	9.8	7.5	1373.0	90.3	1654.1	69.6	1654.1	69.6
103	yl-pk, round	54	31	1.7	2.0	22.8	0.1	2.9	0.1	9.7	22.6	860.2	23.7	1686.1	208.9	1686.1	208.9

[1] A lower case c and r after the analysis number refers to the core and rim of the zircon respectively.

[2] Abbreviations: yllw = yellow; yl-pk= yellowish pink; ac facet = acicular, faceted

[3] Apparent ages are $^{207}Pb/^{206}Pb$ ages calculated using the ^{204}Pb-method of common Pb correction for zircons >1200 Ma, and $^{206}Pb/^{238}U$ ages calculated using the ^{207}Pb-method.

Decay constants: 235U=9.8485x10-10. 238U=1.55125x10-10, 238U/235U=137.88.

Uncertainties include only random errors, and are shown at the 1-sigma level.

Isotope ratios are corrected for Pb/U fractionation by comparison with standard zircon with an age of 545 ± 4 Ma.

Initial Pb composition interpreted from Stacey and Kramers (1975), with uncertainties of 1.0 for 206Pb/204Pb and 0.3 for 207Pb/204Pb.

Appendix C. $^{40}Ar/^{39}Ar$ ages from the Appalachians

Location	Age (Ma)	Phase[1]	Reference	Notes:
New England				
Falmouth-Brunswick seq	266-295	hbl	West et al. (1993)	S. central/W. Maine
	246-249	wh mi	West et al. (1993)	S. central/W. Maine
	241-245	bt	West et al. (1993)	S. central/W. Maine
	226-227	ksp	West et al. (1993)	S. central/W. Maine
Casco Bay Group	333-377	hbl	West et al. (1993)	S. central/W. Maine
	298-307	wh mi	West et al. (1993)	S. central/W. Maine
	288-299	bt	West et al. (1993)	S. central/W. Maine
	265-269	ksp	West et al. (1993)	S. central/W. Maine
Sebago batholith	248	wh mi	West et al. (1993)	S. central/W. Maine
	245-246	bt	West et al. (1993)	S. central/W. Maine
	227	ksp	West et al. (1993)	S. central/W. Maine
Mooselookmeguntic Batholith	358	hbl	Lux and Guidotti (1985)	W. Maine, cent.batho.
	330	hbl	Lux and Guidotti (1985)	W. Maine, SW edge of bath.
Songo Pluton	305-309	hbl	Lux and Guidotti (1985)	W. Maine
western New Hampshire	407-280	hbl	Spear and Harrison (1989)	W. NH, younger to N
	250-315	wh mi	Harrison et al. (1989)	W. NH, younger to W.
eastern Vermont	1026-355	hbl	Spear and Harrison (1989)	S.E. VT
	345	wh mi	Harrison et al. (1989)	S.E. VT
Eastern Green Mountains, VT	470-350	hbl	Sutter and Dallmeyer (1982)	
	450-315	bt	Sutter and Dallmeyer (1982)	
Western Green Mountains, VT	1090-880	hbl	Sutter and Dallmeyer (1982)	
	850-700	bt	Sutter and Dallmeyer (1982)	
northern Massachusetts	420-287	hbl	Spear and Harrison (1989)	N-central MA
	345	wh mi	Harrison et al. (1989)	N. MA
Bronson Hill Anticlinorium	294-249	hbl	Wintsch et al. (2003)	central MA and CT
	235-262	wh mi	Wintsch et al. (2003)	central MA and CT
southern Connecticut Valley	318-254	wh mi	Moecher et al. (1997)	S-SW CT
	374-324	hbl	Moecher et al. (1997)	S-SW CT
Putnam belt	352-273	hbl	Wintsch et al. (1992)	E and central CT
Hope Valley terrane	255-263	hbl	Wintsch et al. (1992)	E and central CT
Esmond Dedham terrane	276-263	hbl	Wintsch et al. (1992)	E and central CT
eastern Reading Prong	470-350	hbl	Sutter and Dallmeyer (1982)	NY and CT
	450-315	bt	Sutter and Dallmeyer (1982)	NY and CT
central Appalachians				
eastern Reading Prong	869-949	hbl	Dallmeyer et al. (1975)	N. NJ and S. NY
	768-819	bt	Dallmeyer et al. (1975)	N. NJ and S. NY
W and S Reading Prong	1090-880	hbl	Sutter and Dallmeyer (1982)	NY, NJ, and PA
	850-700	bt	Sutter and Dallmeyer (1982)	NY, NJ, and PA
Honey Brook Uplands	880	hbl	Sutter et al. (1980)	PA
	850-700	bt	Sutter and Dallmeyer (1982)	PA
Glenarm terrane (W.Chester	375	hbl	Sutter et al. (1980)	PA
Prong)	360	bt	Sutter et al. (1980)	PA

Appendix C (cont.). $^{40}Ar/^{39}Ar$ ages from the Appalachians

Location	Age (Ma)	Phase	Reference	Notes:
central Appalachians (cont.)				
Rosemont Fault Zone	465	hbl	Sutter et al. (1980)	w/I Wissahickon Schist
	410	bt	Sutter et al. (1980)	w/I Wissahickon Schist
Piedmont of SE/SC PA	360	K-Ar, mu, bt	Lapham and Bassett (1964)	Safe Harbor
Piedmont of northern Md	330	K-Ar, mu	Lapham and Bassett (1964)	northern Md Piedmont
Westminster terrane	338-367	wh mi	Krol et al. (1999)	
Potomac terrane	301-310	wh mi	Krol et al. (1999)	
Blue Ridge Anticlinorium, VA	978-956	hbl	Kunk and Burton (1999)	Harper's Ferry, WV
	337-347	wh mi	Kunk and Burton (1999)	Harper's Ferry, WV
	1004	hbl	Kunk and Burton (1999)	Middlesburg, VA
	920	hbl	Kunk and Burton (1999)	Purcellville, VA
	1008	hbl	Kunk and Burton (1999)	Purcellville, VA
	523.9	hbl	Kunk and Burton (1999)	Flint Hill, VA
	313-338	wh mi	Kunk and Burton (1999)	Point of Rocks, MD-VA
	323-347	wh mi	Kunk and Burton (1999)	Point of Rocks, MD-VA
	311-360	wh mi	Kunk and Burton (1999)	Point of Rocks, MD-VA
	338-346	wh mi	Kunk and Burton (1999)	Lovingston, VA
	297-337	wh mi	Kunk and Burton (1999)	Thoroughfare Gap, VA
Southern Appalachians				
northern Blue Ridge	470-350	hbl	Sutter and Dallmeyer (1982)	VA
	450-315	bt	Sutter and Dallmeyer (1982)	VA
Chopawamsic Terrane	317.8	hbl	Burton et al. (2000)	20 km NW of Farmville, VA
Milton belt	313	hbl	Kunk et al. (1995)	E. edge of Milton Belt
	323	hbl	Kunk et al. (1995)	core of Milton Belt
northern Carolina terrane, VA	283.8	hbl	Burton et al. (2000)	35 km NE of Farmville, VA
	260	wh mi	Burton et al. (2000)	7 km south of hbl sample
	274	bt	Burton et al. (2000)	8 km south of hbl sample
	304.4	wh mi	Burton et al. (2000)	N of Chase City, VA
southern Blue Ridge	470-350	hbl	Sutter and Dallmeyer (1982)	NC and VA
	450-315	bt	Sutter and Dallmeyer (1982)	NC and VA
Raleigh Metamorphic Belt	321	hbl	Kunk et al. (1995)	N-cent. NC, near VA
	270	wh mi	Kunk et al. (1995)	N-cent. NC, near VA
	265	bt	Kunk et al. (1995)	N-cent. NC, near VA
Blue Ridge Horses, NC/TN	382-992	hbl	Goldberg and Dallmeyer (1997)	Beech Mtn Thrust Sheet
	518-774	bi	Dallmeyer (1975)	Beech Mtn Thrust Sheet
	366-406	hbl	Goldberg and Dallmeyer (1997)	Pumpkin Patch Thrust Sheet
	327-336	wh mi	Goldberg and Dallmeyer (1997)	Pumpkin Patch Thrust Sheet
	361-398	hbl	Goldberg and Dallmeyer (1997)	Spruce Pine Thrust Sheet
	323-336	wh mi	Goldberg and Dallmeyer (1997)	Spruce Pine Thrust Sheet
Murphy belt, WBR	275-377	mu-bt	Connelly and Dallmeyer (1993)	W. NC and E TN
	420-436	hbl	Connelly and Dallmeyer (1993)	W. NC and E TN
western Blue Ridge	1090-880	hbl	Sutter and Dallmeyer (1982)	TN and NC
	850-700	bt	Sutter and Dallmeyer (1982)	TN and NC
Charlotte Belt	406	hbl	Sutter et al. (1983)	Farm. Gbr., Davie Co., NC
	408	hbl	Sutter et al. (1983)	Wed. Gbr., Union Co., NC
	430	hbl	Sutter et al. (1984)	near SC/NC border

Appendix C (cont.). $^{40}Ar/^{39}Ar$ ages from the Appalachians

Location	Age (Ma)	Phase	Reference	Notes:
Southern Appalachians (cont.)				
Charlotte Belt (cont.)	425-435	wr	Sutter et al. (1983)	count. rk surr. gbrs
	341-278	hbl	Dallmeyer et al. (1986)	north-central SC
	243-292	bt	Dallmeyer et al. (1986)	north-central SC
Six-mile Nappe/Alto Allochthon	332-414	hbl	Dallmeyer (1988)	within Alto Allochthon
	305-312	wh mi	Dallmeyer (1988)	within Alto Allochthon
Chauga Walhalla complex	322-362	hbl	Dallmeyer (1988)	E edge of Alto allocth.
	301-308	wh mi	Dallmeyer (1988)	E edge of Alto allocth.
Carolina slate belt, NC	455	wr	Noel et al. (1988)	central NC, S of Asheboro
	600-586	hbl	Kunk et al. (1995)	W. side of CSB
King's Mountain	318-323	hbl	Horton et al. (1987)	SC/NC border, King's Mtn
	355.9	hbl	Horton et al. (1987)	imm. W of King's Mtn
	421	hbl	Horton et al. (1987)	imm. W of York pluton
	322	hbl	Sutter et al.(1984)	w/I King's Mountain belt
	289-313	bt	Dallmeyer et al. (1986)	S. edge of King's Mountain
	302	hbl	Dallmeyer et al. (1986)	S. edge of King's Mountain
	284.9	wh mi	Horton et al.(1987)	High Shoals granite
NW of Tallulah Falls	332	hbl	Dallmeyer (1988)	northwest of Tallulah Falls
Tallulah Falls dome	322-328	hbl	Dallmeyer (1988)	northwest of Alto
	311-317	wh mi	Dallmeyer (1988)	northwest of Alto
GA Piedmont	241-317	Bt	Dallmeyer (1988)	w/I I. Pied., near SC line
	267-355	Hbl	Dallmeyer (1988)	w/I I. Pied., near SC line
Inner Piedmont	335	hbl	Sutter et al. (1984)	near King's Mountain bdy
	296-300	hbl	Dallmeyer et al. (1986)	SW edge of King's M. B.
	259-271	bt	Dallmeyer et al. (1986)	SW edge of King's M. B.
Clouds Creek granite	268	bt	Dallmeyer et al. (1986)	cent.SC; NW edge of Kiok.
	285-343	bt	Dallmeyer et al. (1986)	cent.SC; NW edge of Kiok.
Southern Carolina Slate Belt	280-331	wr	Dallmeyer et al. (1986)	cent. SC, btwn Kiok. & Char.
Augusta terrane	310-320	wr	Maher et al. (1994)	near coastal plain, GA/SC
Savannah River terrane	272-277	wh mi	Maher et al. (1994)	near coastal plain, GA/SC
southwestern Blue Ridge	1090-880	hbl	Sutter and Dallmeyer (1982)	GA
	850-700	bt	Sutter and Dallmeyer (1982)	GA
Kiokee Belt	288	hbl	Sacks et al. (1989)	near Augusta Georgia
	288-324	hbl	Dallmeyer et al. (1986)	N. half Kiokee Belt, cent. SC
	278-288	bt	Dallmeyer et al. (1986)	N. half Kiokee Belt, cent. SC
eastern Blue Ridge, GA	354-362	hbl	Dallmeyer (1989)	GA; at Brasstown Bald
	322-324	wh mi	Dallmeyer (1989)	GA; at Brasstown Bald
	323	bt	Dallmeyer (1989)	GA; at Brasstown Bald
Uchee Belt	285	wh mi	Steltenpohl et al. (1992)	near coastal plain, AL/GA
	288-297	hbl	Steltenpohl et al. (1992)	near coastal plain, AL/GA
Pine Mountain Group	283-286	wh mi	Steltenpohl et al. (1992)	N. of BFFZ, S. of Towaliga
southern Talledega slate belt	387-417	K/Ar wr	Kish (1990)	central AL
Wiggins uplift	307-315	wh mi	Dallmeyer (1989)	S.W. AL/ S.E. MS
	309	hbl	Dallmeyer (1989)	S.W. AL/ S.E. MS
Subsurface Piedmont	292	bt	Dallmeyer (1989)	S.W. AL

[1] Abbreviations of mineral phases: bt=biotite, hbl=hornblende, wh mi=white mica (phengite+muscovite), ksp=alkali feldspar, wr=whole rock(phyllite), mu=muscovite

REFERENCES CITED

Adams, M.G., and Su, Q., 1996, The nature and timing of deformation in the Beech Mountain thrust sheet between Grandfather Mountain and Mountain City windows in the Blue Ridge of northwestern North Carolina: Journal of Geology, v. 104, p. 197-213.

Albarede, F., and Michard, A., 1987, Evidence for slowly changing $^{87}Sr/^{86}Sr$ in runoff from freshwater limestones of southern France: Earth and Planetary Science Letters, v. 64, p. 55-65.

Albarede, F., Michard, A., Minster, J.F., and Michard, G., 1981, $^{87}Sr/^{86}Sr$ ratios in hydrothermal waters and deposits from the East Pacific Rise at 21 degrees N.: Earth and Planetary Science Letters, v. 55, p. 229-236.

Aleinikoff, J.N., Horton, J.W., Jr., Drake, A.A., Jr., and Fanning, C.M., 2002, SHRIMP and conventional U-Pb ages of Ordovician granites and tonalities in the central Appalachian Piedmont: implications for Paleozoic tectonic events: American Journal of Science, v. 302, p. 50-75.

Aleinikoff, J.N., Zartman, R.E., Walter, M., Rankin, D.W., Lyttle, P.T., and Burton, W.C., 1995, U-Pb ages of metarhyolites of the Catoctin and Mount Rogers Formations, central and southern Appalachians: evidence for two pulses of Iapetan rifting: American Journal of Science, v. 295, p.428-454.

Anderson, J.L., 1983, Proterozoic anorogenic granite plutonism of North America, *in* Medaris Jr., L.G., Byers, C.W., Shanks, W.C., eds., Proterozoic Geology: Geological Society of America Memoir, v. 302, p. 50-75.

Angiolini, L., Balini, M., Garzanti, E., Nicora, A., and Tintori, A., 2003, Gondwanan deglaciation and opening of Neotethys: the Al Khlata and Saiwan Formations of Interior Oman: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 196, p. 99-123.

Arkle, T., Jr., Beissel, D.R., Larese, R.E., Nuhfer, E.B., Patchen, D.G., Smosna, R.A., Gillespie, W.H., Lund, R., Norton, C.W., and Pfefferkorn, H.W., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States-West Virginia and Maryland: U.S. Geological Survey Professional Paper, v. 1110-D, p. 1-35.

Aronson, J.L., and Lewis, T., 1994, Ages of detrital white mica from Devonian-Pennsylvanian strata of the north central Appalachian Basin: Dominance of the Acadian orogen as provenance: Journal of Geology, v. 102, p. 685-696.

Aronson, J.L., Farrington, S.A., 2000, Origin of 2M and 1M illite in the Devonian shales of the Appalachian basin. Geological Society of America-Abstracts with Programs, v. 32, n.7, p. 366.

Berner, R.A., 2001, Modeling atmospheric O_2 over Phanerozoic time. Geochimica Cosmochimica Acta, v. 65, p.685-694.

Berner, R.A., Beerling, D.J., Dudley, R., Robinson, J.M., and Wildman, R.A., Jr., 2003, Phanerozoic atmospheric oxygen: Annual Review of Earth and Planetary Sciences, v. 31, p. 105-134.

Berner, R.A., Petsch, S.T., Lake, J.A., Berrling, D.J., Popp, B.N., Lane, R.S., Westley, M.B., Cassar, N., Woodward, F.I., and Quick, W.P., 2000, Isotope fractionation and atmospheric oxygen; implications for Phanerozoic O_2 evolution: Science, v. 287, p. 1630-1633.

Bethke, C.M., Marshak, S., 1990, Brine migrations across North America--the plate tectonics of groundwater: Annual Reviews of Earth and Planetary Science, v. 18, p. 287-315.

Bickford, M.E., Van Schmus, W.R., and Zietz, I.,1986, Proterozoic history of the Midcontinent region of North America: Geology, v. 14, p. 492-496.

Bickford, M.E., Soegaard, K., Nielson, K.C., and McLelland, J.M., 2000, Geology and geochronology of Grenville-age rocks in the Van Horn and Franklin Mountains area, West Texas; implications for the tectonic evolution of Laurentia during the Grenville: Geological Society of America Bulletin, v. 112, p. 1134-1148.

Bird, J.M. and Dewey, J.F.,1970, Lithosphere plate-continental margin tectonics and the evolution of the Appalachian orogen: Geological Society of America Bulletin, v. 81, p. 1031-1059.

Blum, J.D. and Erel, Y.,1997, Rb-Sr isotope systematics of a granitic soil chronosequence: The importance of biotite weathering: Geochimica Cosmochimica Acta, v. 61, p. 3192-3204.

Bosbyshell, H., Sinha, A.K., Crawford, M.L., Fleming, P., Srogi, L., and Lutz, T.M., 1998, Thermal evolution of a convergent orogen: new U/ Pb ages of monazite and zircon from the central Appalachian Piedmont: Geological Society of America-Abstracts with Programs, v. 30, n. 7, p. 125.

Bowring, S.A., 1984, U-Pb zircon ages of granitic boulders in the Ordovician Blakely Sandstone, Arkansas and implications for their provenance. *In* Stone, C.G. and Haley, B.R., eds, A guidebook to the geology of the central and southern Ouachita Mountains Arkansas: Arkansas Geological Community Guidebook, v.84, n.2, p. 123.

Bradley, D.C., 1982, Subsidence in late Paleozoic basins in the northern Appalachians: Tectonics, v.1, p.107-123.

Brass, G.W., 1976, The variation of marine $^{87}Sr/^{86}Sr$ ratio during Phanerozoic time: Interpretation using a flux model: Geochimica Cosmochimica Acta, v. 40, p. 721-730.

Bream, B.R., 2002, The southern Appalachian Inner Piedmont: New perspectives based on recent detailed geologic mapping, Nd isotopic evidence, and zircon geochronology, *In* Hatcher, R.D., Jr., and Bream, B.R., eds., Inner Piedmont geology in the South Mountains-Blue Ridge Foothills and the southwestern Brushy Mountains, central-western North Carolina: North Carolina Geological Survey, Carolina Geological Society Annual Field Trip Guidebook, p. 45-63.

Bream, B.R., Hatcher, R.D., Jr., Miller, C.F., and Fullagar, P.D., 2004, Detrital zircon ages and Nd isotopic data from the southern Appalachian crystalline core, Georgia, South Carolina, North Carolina, and Tennessee: New provenance constraints for part of the Laurentian margin, *In* Tollo, R.P., Corriveau, L., McClelland, J., and Bartholomew, M.J., eds., Proterozoic tectonic evolution of the Grenville orogen in North America: Geological Society of America Memoir 197, p. 459-475.

Brown, M., and Solar, G., 1999, The mechanism for ascent and emplacement of granite magma during transpression: Tectonophysics, v. 312, p. 1-33.

Bryant, J.D., Jones, D.S., and Mueller, P.A., 1995, Influence of freshwater flux $^{87}Sr/^{86}Sr$ chronostratigraphic in marginal marine environments and dating of vertebrate and invertebrate faunas: Journal of Paleontology, v. 69, p. 1-6.

Bullard, E.C., Everett, J.E., and Smith, A.G., 1965, The fit of the continents around the Atlantic: Royal Society of London Philosophical Transactions, v. A258, p. 41-51.

Burke, W.H. and Hetherington, E.A., 1984, Normalized $^{87}Sr/^{86}Sr$ by multiple collection and comparison to a standard: Chemical Geology, v. 46, p. 265-268.

Burke, W.H., Denison, R.E., Hetherington, E.A., Koepnick, R.B., Nelson, H.F., and Otto, J.B., 1982, Variation of seawater $^{87}Sr/^{86}Sr$ throughout Phanerozoic time: Geology, v. 10, p. 516-519.

Burton, W.C., Kunk, M.J., and Marr, J.D., 2000, $^{40}Ar/^{39}Ar$ age constraints on the timing of Alleghanian metamorphism in the central and southern Virginia Piedmont. Abstracts with Programs-Geological Society of America, v. 32 (2), p. 8.

Butts, C., 1926, Geology of Alabama. Alabama Geological Survey Special Report 14, p. 41-230.

Caputo, M.V., and Crowell, J.C., 1985, Migration of glacial centers across Gondwana during Paleozoic Era: Geological Society of America Bulletin, v. 96, p. 1020-1036.

Carroll, A.R., Wilkinson, B.H., and Owen, R.M., 1983, Lawrence Lake, Michigan; Holocene carbonate deposition in a temperate region lacustrine system. Geological Society of America Abstracts with Programs, v. 15-6, p. 539.

Cassle, C.F., Gierlowski-Kordesch, E.H., and Martino, R.L., 2003, Late Pennsylvanian carbonates of the northern Appalachian basin: criteria to distinguish brackish and freshwater conditions. Geological Society of America Abstracts with Programs, v. 35 (6), p. 600.

Cawood, P.A., and Nemchin, A.A., 2001, Paleogeographic development of the east Laurentian margin: Constraints from U-Pb dating of detrital zircons in the Newfoundland Appalachians: Geological Society of America Bulletin, v. 113, p. 1234-1246.

Cawood, P.A., McCausland, P.J.A., and Dunning, G.R., 2001, Opening Iapetus: constraints from the Laurentian margin in Newfoundland: Geological Society of America Bulletin, v. 113, p. 443-453.

Cecil, C.B., 1990, Paleoclimate controls on stratigraphic repetition of chemical and siliciclastic rocks: Geology, v. 18, p. 533-536.

Coffield, D.Q., 1993, Evidence for wrenching during mid-Permian extension in central East Greenland, in Vorren, T.O.; Bergsager, E.; Dahl, S.O.A.; Holter, E.; Johansen, B.; Lie, E.; Lund, T.B., eds., Arctic geology and petroleum: Proceedings of the Norwegian Petroleum Society conference. Norwegian Petroleum Society Special Publication. v. 2, p. 89-98.

Coler, D.G., and Samson, S.D., 2000, Characterization of the Spring Hope and Roanoke Rapids terranes, southern Appalachians; a U-Pb geochronologic and Nd isotopic study: Geological Society of America Abstracts with Programs, v. 32 (1), p. 11-12.

Coler, D.G., Samson, S.D., and Speer, J.A., 1997, Nd and Sr isotopic constraints on the source of Alleghanian granites in the Raleigh metamorphic belt and Eastern Slate belt, southern Appalachians, U.S.A.: Chemical Geology, v. 134, p. 257-275.

Coler, D.G., Wortman,G.L., Samson, S.D., Hibbard, J.P. and Stern, R., 2000, U-Pb geochronologic, Nd isotopic, and geochemical evidence for the correlation of the Chopawamsic and Milton terranes, Piedmont Zone, Southern Appalachian Orogen: Journal of Geology, v. 108, n.4, p. 363-380.

Collins, H.R., and Smith, B.E., 1977, Geology and mineral resources of Washington County, Ohio: Ohio Division of Geological Survey, Bulletin 66, 134 p.

Collinson, C., Sargent, M.C., and Jennings, J.R., 1988, Eastern Interior Basin., *In* Sloss, L.L., ed., Sedimentary Cover-North American Craton, U.S.: Boulder, Colorado, Geological Society of America, The Geology of North America, v. D-2, p. 383-426.

Colton, G.W., 1970, The Valley and Ridge and Appalachian Plateau; stratigraphy and sedimentation; the Appalachian Basin; its depositional sequences and their geologic relationships, *In* Fisher. G.W., Pettijohn, F.J., Reed, J.C., Jr., and Weaver, K.N., eds., Studies of Appalachian geology, central and southern: Interscience Publishing, New York, New York, p. 5-47

Connelly, J.B., and Dallmeyer, R.D., 1993, Polymetamorphic evolution of the western Blue Ridge; evidence from $^{40}Ar/^{39}Ar$ whole-rock slate/phyllite and muscovite ages: American Journal of Science, v. 293, p. 322-359.

Cook, F.A., Brown,L.D., Kaufman, S., Oliver, J.E., and Petersen, T.A., 1981, COCORP seismic profiling of the Appalachian orogen beneath the Coastal Plain of Georgia: Geological Society of America Bulletin, v. 92, p. 738-748.

Dallmeyer, R.D., 1989, Late Paleozoic thermal evolution of crystalline terranes within portions of the U.S. Appalachian orogen, *In* Hatcher, R.D., Jr., Thomas, W.A. and Viele, G.W., eds., The Appalachian-Ouachita Orogen in the United States: Boulder, Colorado, Geological Society of America, The Geology of North America, v. F-2, p.417-444.

Dallmeyer, R.D., Wright, J.E., Secor, D.T., Jr., and Snoke, A.W., 1986, Character of the Alleghanian orogeny in the southern Appalachians: Part II. Geochronological constraints on the tectonothermal evolution of the eastern Piedmont in South Carolina: Geological Society of America Bulletin, v. 97, p. 1329-1344.

Dallmeyer, R.D., 1988, Late Paleozoic tectonothermal evolution of the western Piedmont and eastern Blue Ridge, Georgia: controls on the chronology of terrane accretion and transport in the southern Appalachian orogen: Geological Society of America Bulletin, v. 100, p. 702-713.

Dallmeyer, R.D., Sutter, J.F., and Baker, D.J., 1975, Incremental $^{40}Ar/^{39}Ar$ ages of biotite and hornblende from the northeastern Reading Prong; their bearing on late Proterozoic thermal and tectonic history: Geological Society of America Bulletin, v. 86, p. 1435-1443.

Dalziel, I.W.D., 1997, Neoproterozoic-Paleozoic geography and tectonics: review, hypothesis, environmental speculation: Geological Society of America Bulletin v. 109, p. 16-42.

Davis, M.W., and Ehrlich, R., 1974, Late Paleozoic crustal composition and dynamics in the southeastern United States, *In* Briggs, G., ed., Carboniferous of the southeastern United States: Geological Society of America Special Paper 148, p. 171-185.

Davis, D., Suppe, J., and Dahlen, F.A., 1983, Mechanics of fold-and-thrust belts and accretionary wedges: Journal of Geophysical Research, v. B-88, p. 1153-1172.

Dean, S.L., Kulander, B.R., and Skinner, J.M., 1988, Structural chronology of the Alleghanian orogeny in southeastern West Virginia: Geological Society of America Bulletin, v. 100, p. 299-310.

Denison, R.E., Koepnick, R.B., Burke, W.H., Hetherington, E.A., and Fletcher, A., 1994, Construction of the Mississippian, Pennsylvanian, and Permian seawater $^{87}Sr/^{86}Sr$ curve: Chemical Geology, v. 112, p. 145-167.

Denison, R.E., Koepnick, R.B., Burke, W.H., and Hetherington, E.A., 1998, Construction of the Cambrian and Ordovician seawater $^{87}Sr/^{86}Sr$ curve: Chemical Geology, v. 152, p. 325-340.

Dennis, A.J., and Wright, J.E., 1995, Mississippian (ca. 326-323 Ma) U-Pb crystallization ages for two granitoids in Spartanburg and Union Counties, South Carolina: Southeastern Geology, v. 38, p.23-28.

Dickinson, W.R., Beard, L.S., Brakenridge, G.R., Erjavec, J.L., Ferguson, R.C., Inman, K.F., Knepp, R.A., Lindberg, F.A., and Ryberg, P.T., 1983, Provenance of North American Phanerozoic sandstones in relation to tectonic setting: Geological Society of America Bulletin, v. 94, p. 222-235.

Dickinson, W.R., and Suzcek, C.A., 1979, Plate tectonics and sandstone compositions: American Association of Petroleum Geologists, v. 63, p. 2164-2182.

DiMichele, W.A., Pfefferkorn, H.W., and Gastaldo, R.A., 2001, Response of late Carboniferous and early Permian plant communities to climate change: Annual Reviews of Earth and Planetary Sciences, v.29, p.461-487.

Dodson, M.H., 1973, Closure temperature in cooling geochronological and petrological systems: Contributions to Mineralogy and Petrology, v. 40, p. 259-274.

Dominic, D.F., 1988, Quantitative interpretation of Upper Pennsylvanian-Lower Permian sandstones of northern West Virginia using mathematical models of sedimentation: Ph.D. dissertation (unpub.), West Virginia University, 151 p.

Donaldson, A.C., Renton, J.J., and Presley, M.W., 1985, Pennsylvanian deposystems and paleoclimates of the Appalachians: International Journal of Coal Geology, v. 5, p. 167-193.

Dorais, M.J., Wintsch, R.P., and Becker, H., 2001, The Massabesic gneiss complex, New Hampshire: A study of a portion of the Avalon terrane: American Journal of Science, v. 301, p. 657-682.

Druhan, R., Butler, J., Horton, J.W., Jr., and Stoddard, E., 1988, Tectonic significance of the Nutbrush Creek fault zone, eastern Piedmont of North Carolina and Virginia: Geological Society of America Abstracts with Programs, v. 20, n.4, p. 261.

Drummond, C.N., Wilkinson, B.H., Lohmann, K.C., and Smith, G.R., 1993, Effect of regional topography and hydrology on the lacustrine isotopic record of Miocene paleoclimate in the Rocky Mountains: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 101, p. 67-79.

Duston, N.M., Owen, R.M., and Wilkinson, B.H., 1986, Water chemistry and sedimentological observations in Littlefield Lake, Michigan: Implications for lacustrine marl deposition: Environmental Geology and Water Sciences, v. 8, p. 229-236.

Edelman, S.H., Liu, A., and Hatcher, R.D., Jr., 1987, The Brevard zone in South Carolina and adjacent areas: An Alleghanian orogen-scale dextral shear zone reactivated as a thrust fault: Journal of Geology, v. 95, p. 793-806.

Edmunds, W.E., 1992, Early Pennsylvanian (middle Morrowan) marine transgression in south-central Pennsylvania: Northeastern Geology, v. 14, p. 225-231.

Edmunds, W.E., 1988, The Pottsville Formation of the anthracite region, *In* Inners, J.D., ed., Geology of the Wyoming-Lackawanna Valley and its mountain rim, northeastern Pennsylvania: Harrisburg, Pennsylvania, Guidebook for the Annual Field Conference of Pennsylvania Geologists, 40-50.

Edmunds, W.E., Berg, T.M., Sevon, W.D., Piotrowski, R.C., Heyman, L., and Rickard, L.V., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States-Pennsylvania and New York: U.S. Geological Survey Professional Paper 1110-B, p. 1-33.

Edmunds, W.E., Skema, V.W., and Flint, N.K., 1999, Pennsylvanian, *In* Shultz, C.H., ed., The Geology of Pennsylvania. Pennsylvania Geological Survey, Harrisburg, PA, p. 149-169.

Edmunds, W.E., 1999, Pennsylvanian-Permian transition and Permian, *In* Shultz, C.H., ed., The Geology of Pennsylvania, Pennsylvania Geological Survey, Harrisburg, PA, p. 171-177.

Eggleston, J.R., and Dean, W.E., 1976, Freshwater stromatolitic bioherms in Green Lake, New York, *In* Walter, M.R., ed., Stromatolites. Elsevier, Amsterdam, p. 479-488.

Elliott, W.C., and Aronson, J.L., 1987, Alleghanian episode of K-bentonite illitization in the southern Appalachian basin: Geology v. 15, p.735-739.

Elliott, W.C., and Aronson, J.L., 1993, The timing and extent of illite formation in Ordovician K-bentonites at the Cincinnati arch, the Nashville dome and north-eastern Illinois basin: Basin Research, v.5, p.125-135.

Engelder, T., 2004, Tectonic implications drawn from differences in the surface morphology on two joint sets in the Appalachian Valley and Ridge, Virginia: Geology, v. 32, p. 413-416.

Englund, K.J., 1974, Sandstone distribution patterns in the Pocahontas Formation of southwest Virginia and southern West Virginia. *In* Briggs, G., ed., Carboniferous of the southeastern United States: Geological Society of America Special Paper, v. 148, p. 31-45.

Englund, K.J., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States-Virginia: U.S. Geological Survey Professional Paper 1110-C, p.1-21.

Englund, K.J., and DeLaney, A.O., 1966, Intertonguing relations of the Lee Formation in southwestern Virginia: U.S. Geological Survey Professional Paper 550-D, p. D47-D52.

Eriksson, K.A., Campbell, I.H., Palin, J.M., Allen, C.M., and Bock, B., 2004, Evidence for multiple recycling in Neoproterozoic through Pennsylvanian sedimentary rocks of the central Appalachian basin: Journal of Geology, v. 112, p. 261-276.

Eusden, J.D., Jr., Guzofski, C.A., Robinson, A.C., and Tucker, R.D., 2000, Timing of the Acadian orogeny in northern New Hampshire: Journal of Geology, v. 108, p.219-232.

Faill, R.T., 1998, A geologic history of the north-central Appalachians, part 3. The Alleghany orogeny: American Journal of Science, v. 298, p. 131-179.

Ferm, J.C., 1974, Carboniferous environmental models in eastern United States and their significance, *In* Briggs, G., ed., Carboniferous of the southeastern United States: Geological Society of America Special Paper 148, p. 79-95.

Ferrill, B.A., and Thomas, W.A., 1988, Acadian dextral transpression and synorogenic sedimentary successions in the Appalachians: Geology, v. 16, p. 604-608.

Francheteau, J., and LePinchon, X., 1972, Marginal fracture zones as structural framework of continental margins in South Atlantic Ocean: American Association of Petroleum Geologists Bulletin, v. 56, p. 991-1007.

Fullagar, P.D., 1971, Age and origin of plutonic intrusions in the Piedmont of the southeastern Appalachians: Geological Society of America Bulletin, v. 82, p. 2845-2862.

Fullagar, P.D., and Butler, J.R., 1979, 325 to 265 m.y.-old granitic plutons in the Piedmont of the southeastern Appalachians: American Journal of Science, v. 279, p. 161-185.

Gates, A.E., Simpson, C., and Glover, L. III, 1986, Appalachian Carboniferous dextral strike-slip faults: An example from Brookneal, Virginia: Tectonics, v. 5, n.1, p. 119-133.

Gates, A.E., Speer, J.A., and Pratt, T.L., 1988, The Alleghanian southern Appalachian Piedmont: A transpressional model: Tectonics, v. 7, p. 1307-1324.

Gates, A.E., and Volkert, R.A., 2004, Vestiges of an Iapetan rift basin in the New Jersey Highlands: Implications for the Neoproterozoic Laurentian margin: Journal of Geodynamics, v. 37, p. 381-409.

Gaudette, H.E., Fairbairn, H.W., Kovach, A., and Hussey, A.M. II, 1975, Preliminary Rb-Sr whole rock age determinations of granitic rocks in southwestern Maine: Geological Society of America Abstracts with Programs, v. 7 (1), p. 62-63.

Getty, S.R., and Gromet, L.P., 1992a, Evidence for extension at the Willimantic dome, Connecticut; implications for the Late Paleozoic tectonic evolution of the New England Appalachians: American Journal of Science, v. 292, p. 398-420.

Getty, S.R., and Gromet, L.P., 1992b, Geochronological constraints on ductile deformation, crustal extension, and doming about a basement-cover boundary, New England Appalachians: American Journal of Science, v. 292, p. 359-397.

Glascock, J.M., and Gierlowski-Kordesch, E.H., 2002, Sedimentologic analysis of the Benwood limestone (Monongahela Group) in Morgan County, Ohio: Geological Society of America Abstracts with Programs, v. 34 (6), p.383.

Gleason, P.J., 1972, The origin, sedimentation, and stratigraphy of a calcic mud located in the southern freshwater Everglades [M.S. thesis]: Pennsylvania State University.

Gleason, J.D., Finney, S.C., and Gehrels, G.E., 2002, Paleotectonic implications of a Mid-to Late-Ordovician provenance shift, as recorded in sedimentary strata of the Ouachita and southern Appalachian Mountains: Journal of Geology, v. 110, p. 291-304.

Glover, L., III, Speer, J.A., Russell, G.S., and Farrar, S.S., 1983, Ages of regional metamorphism and ductile deformation in the central and southern Appalachians: Lithos, v. 16, p. 223-245.

Golanka, J., Ross, M.I., and Scotese, C.R., 1994, Phanerozoic paleogeographic and paleoclimatic modeling maps, *In* Embry, A.F.; Beauchamp, B.; and Glass, D.J., eds., Pangea: Global Environments and Resources: Canadian Society of Petroleum Geologists Memoir 17, p. 1-47.

Goldberg, S.A., and Dallmeyer, R.D., 1997, Chronology of Paleozoic metamorphism and deformation in the Blue Ridge thrust complex, North Carolina and Tennessee: American Journal of Science, v. 297, p. 488-526.

Goldberg, S.A., Butler, J.R., and Fullagar, P.D., 1986. The Bakersville dike swarm: Geochronology and petrogenesis of Late Proterozoic basaltic magmatism in the southern Appalachian Blue Ridge: American Journal of Science, v. 286, p. 403-430.

Goldberg, S.A., and Steltenpohl, M.G., 1990, Timing and characteristics of Paleozoic deformation and metamorphism in the Alabama inner Piedmont: American Journal of Science, v. 290, p. 1169-1200.

Goldstein, A.G., and Hepburn, J.C., 1999, Possible correlations of the Norumbega fault system with faults in southeastern New England, *In* Ludman, A. and West, D.P., Jr., eds., Norumbega fault system of the Northern Appalachians: Boulder, Colorado, Geological Society of America Special Paper 331, p. 73-83.

Graham, S.A., Ingersoll, R.V., and Dickinson, W.R., 1976, Common provenance for lithic grains in Carboniferous sandstones from Ouachita Mountains and Black Warrior basin: Journal of Sedimentary Petrology, v. 46, p.620-632.

Gray, M.B., and Zeitler, P.K., 1997, Comparison of clastic wedge provenance in the Appalachian foreland using U/ Pb ages of detrital zircons: Tectonics, v. 16, p. 151-160.

Gromet, L.P., 1989, Avalonian terranes and late Paleozoic tectonism in southeastern New England: constraints and problems: Geological Society of America Special Paper 230, p. 193-211.

Harrison, T.M., Spear, F.S., and Heizler, M.T., 1989, Geochronologic studies in central New England II: Post-Acadian hinged and differential uplift: Geology, v. 17, p. 185-189

Hatcher, R.D., Jr., 1972, Developmental model for the southern Appalachians: Geological Society of America Bulletin, v. 83, p. 2735-2760.

Hatcher, R.D., Jr., 1978, Tectonics of the western Piedmont and Blue Ridge: Review and speculation: American Journal of Science, v. 278, p. 276-304.

Hatcher, R.D., Jr., 1987, Tectonics of the southern and central Appalachian internides: Annual Reviews of Earth and Planetary Sciences, v. 15, p. 337-362.

Hatcher, R.D., Jr., 1989, Tectonic synthesis of the U.S. Appalachians, *In* Hatcher, R.D., Jr., Thomas, W.A. and Viele, G.W., eds., The Appalachian-Ouachita Orogen in the United States: Boulder, Colorado, Geological Society of America, The Geology of North America, v. F-2, p. 233-318.

Hatcher, R.D., Jr., 2002, Alleghanian (Appalachian) orogeny, a product of zipper tectonics: Rotational transpressive continent-continent collision and closing of ancient oceans along irregular margins, *In* Martinez, J.R., Hatcher, R.D., Jr., Arenas, R., and Diaz Garcia, F., eds., Variscan-Appalachian dynamics: The building of the late Paleozoic basement: Geological Society of America Special Paper 364, p. 199-208.

Hatcher, R.D., Jr., 2001, Rheological partitioning during multiple reactivation of the Paleozoic Brevard fault zone, southern Appalachians, USA, *In* Holdsworth, R.E., Strachan, R.A., Magloughlin, J.F., and Knipe, R.J., eds., The nature and tectonic significance of fault zone weakening: Geological Society of London Special Publication186, p. 257-271.

Hatcher, R.D., Jr., Williams, R.T., 1986, Mechanical model for single thrust sheets: Geological Society of America Bulletin, v. 97, p. 975-985.

Hatcher, R.D., Jr., Bream, B.R., Fullagar, P.D., Miller, C.F., Carrigan, C.W., 2002, Hints of the Mid-Continent beyond the Grenville orogen and Gondwanan inheritance in Appalachian terranes: Geological Society of America Abstracts with Programs, v. 34 (6), p. 78.

Hayes, J.M., Strauss, H., and Kaufman, A.J., 1999, The abundance of ^{13}C in marine organic matter and isotopic fractionation in the global biogeochemical cycle of carbon during the past 800 Ma, *In*: Veizer, J., ed., Earth system evolution; geochemical perspective: Chemical Geology, v. 161, n. 1-3, p. 103-125.

Hayward, J.A., and Gaudette, H.E., 1984, Carboniferous age of the Sebago and Effingham plutons, Maine and New Hampshire: Geological Society of America Abstracts with Programs, v. 16 (1), p. 22.

Heatherington, A., 1998, Age and isotopic composition of the Appling Granite, eastern Georgia: Southeastern Geology, v. 38, p. 21-27.

Heatherington, A., and Mueller, P., 1997, Zircon systematics of Alleghanian granites of the southernmost Appalachians; implications for terrane provenance: Geological Society of America Abstracts with Programs, v. 29 (3), p. 23

Heatherington, A.L., Mueller, P.A., and Wooden, J.L., 1999, Alleghanian plutonism in the Suwannee terrane. Geological Society of America Abstracts with Programs, v. 31 (7), p. 117.

Heatherington, A.L., 1998, Age and isotopic composition of the Appling Granite, eastern Georgia: Southeastern Geology, v. 38, p. 21-27.

Heatherington, A.L., and Mueller, P.A., 1999, Lithospheric sources of north Florida, USA, tholeiites and implications for the origin of the Suwannee terrane: Lithos, v. 46, p. 215-233.

Hermes, O.D., and Zartman, R.E., 1985, Late Precambrian and Devonian plutonic terrane within the Avalon Zone of Rhode Island: Geological Society of America Bulletin, v. 96, p. 272-282.

Hermes, O.D., and Zartman, R.E., 1992, Late Proterozoic and Silurian alkaline plutons within the southeastern New England Avalon zone: Journal of Geology, v. 100, p. 477-486.

Hess, H.H., 1939, Island arcs, gravity anomalies, and serpentine intrusions: A contribution to the ophiolite problem: International Geological Congress, Moscow, 1937, v. 2, p. 262-283.

Hibbard, J., 2000, Docking Carolina: Mid-Paleozoic accretion in the southern Appalachians: Geology, v. 28, p. 127-130.

Hibbard, J., Tracy, R.J., and Henika, W.S., 2003, Smith River allochthon: A southern Appalachian peri-Gondwanan terrane emplaced directly on Laurentia?: Geology, v. 31, p. 215-218.

Hibbard, J.P., Shell, G.S., Bradley, P.J., Samson, S.D., and Wortman, G.L., 1998, The Hyco shear zone in North Carolina and southern Virginia: Implications for the Piedmont Zone-Carolina Zone boundary in the southern Appalachians: American Journal of Science, v. 298, p. 85-107.

Hibbard, J.P., Stoddard, E.F., Secor, D.T., Jr., and Dennis, A.J., 2002, The Carolina zone: overview of Neoproterozoic to early Paleozoic peri-Gondwanan terranes along the eastern flank of the southern Appalachians: Earth-Science Reviews, v. 57, p. 299-339.

Hibbard, J.P., Tracy, R.J., and Henika, W.S., 2003, Smith River Allochthon; a Southern Appalachian peri-Gondwanan terrane emplaced directly on Laurentia?: Geology, v. 31, p. 215-218.

Hillaire, M.C., and Casanova, J., 1987, Isotopic hydrology and paleohydrology of the Magadi (Kenya)-Natron (Tanzania) Basin during the late Quaternary: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 58, p. 155-181.

Hoffman, P.F., 1989, Precambrian geology and tectonic history of North America. *In* Bally, A.W. and Palmer, A.R., eds. Geology of North America-an overview (Geology of North America, vol. A). Boulder, Colorado, Geological Society of America, p. 447-512.

Hogan, J.P., and Gilbert, M.C., 1998, The Southern Oklahoma Aulacogen: a Cambrian analog for Mid-Proterozoic AMCG (anorthosite-mangerite-charnockite-granite) complexes? *In* Hogan, J.P. and Gilbert, M.C., eds., Central North America and other regions. Basement Tectonics 12, Dordrecht, Kluwer Academic Publisher, p. 39-78.

Horton, J.W., Jr., Aleinikoff, J.N., Izett, G.A., Naesar, C.W., and Naesar, N.D., 2001, Crystalline rocks from the first corehole to basement in the Chesapeake Bay impact structure, Hampton, Virginia: Geological Society of America-Abstracts with Programs, v. 33(6), p. 448.

Horton, J.W., Jr., Drake, A.A., Jr., Rankin, D.W., 1989, Tectonostratigraphic terranes and their Paleozoic boundaries in the central and southern Appalachians, *In* Dallmeyer, R.D., ed., Terranes in the circum-Atlantic Paleozoic orogens: Geological Society of America Special Paper 230, p. 213-245.

Horton, J.W., Jr., Sutter, J.F., Stern, T.W., and Milton, D.J., 1987, Alleghanian deformation, metamorphism, and granite emplacement in the central Piedmont of the southern Appalachians: American Journal of Science, v. 287, p. 635-660.

Horton, J.W., Jr., Drake, A.A., Jr., Rankin, D.W., 1989, Tectonostratigraphic terranes and their Paleozoic boundaries in the central and southern Appalachians. *In* Dallmeyer, R.D., ed., Terranes in the circum-Atlantic Paleozoic orogens: Geological Society of America Special Paper 230, p. 213-245.

Hoque, M.U., 1968, Sedimentologic and paleocurrent study of Mauch Chunk sandstones (Mississippian), south-central and western Pennsylvania: American Association of Petroleum Geologists Bulletin, v. 52, p. 246-263.

Idleman, B.D., 1996, $^{40}Ar/^{39}Ar$ dating of detrital minerals from western Newfoundland sandstones: insights into provenance and tectonics along the Laurentian margin of Iapetus: Geological Society of America Abstracts with Programs, v. 28(7), p. 439.

Idleman, B.D., 1998, Late Caradocian exhumation of the Taconic orogen in western Newfoundland: evidence from single-crystal $^{40}Ar/^{39}Ar$ dating of detrital muscovite: Geological Association of Canada-Mineralogical Association of Canada Annual Meeting, v. 23, p. 83.

Ingle-Jenkins, S., Mueller, P.A., Heatherington, A.L., Offield, T.W., 1998, Age and origin of the Uwharrie Formation and Albemarle Group, Carolina slate belt implications from U-Pb and Sm-Nd systematics: Geological Society of America Abstracts with Programs, v. 30 (7), p. 239.

Inners, J.D., and Lentz, L.J., 1988, Description of section measured on Interstate Route 81 near Delano, Schuylkill Couty, Pennsylvania. *In* Inners, J.D., ed., Geology of the Wyoming-Lackawanna Valley and its mountain rim, northeastern Pennsylvania: Harrisburg, Pennsylvania, Guidebook for the Annual Field Conference of Pennsylvania Geologists, p. 197-208.

Johnsson, M.J., 1993, The system controlling the composition of clastic sediments. *In* Johnsson, M.J. and Basu, A., eds., Processes controlling the composition of clastic sediments: Geological Society of America Special Paper 284, p. 1-19.

Jordan, T.E., 1981, Thrust loads and foreland basin evolution, Cretaceous, western United States: American Association of Petroleum Geologists Bulletin, v. 65, p. 2506-2520.

Jordan, T.E., 1995, Retroarc foreland and related basins. *In* Busby, C.J., and Ingersoll, R.V., eds., Tectonics of sedimentary basins. Cambridge, Mass., Blackwell Science, p. 331-362.

Karabinos, P., Samson, S.D., Hepburn, J.C., and Stoll, H.M., 1998, Taconian orogeny in the New England Appalachians: collision between Laurentia and the Shelburne Falls arc: Geology, v. 26, p. 215-218.

Keith, M.L., and Weber, J.N., 1964, Carbon and oxygen composition of selected limestones and fossils: Geochimica Cosmochimica Acta, v. 28, p. 1787-1816.

Keller, G.R., Braile, L.W., McMechan, G.A., Thomas, W.A., Harder, S.H., Chang, W.F., Jardine, W.G., 1989, Paleozoic continent-ocean transition in the Ouachita Mountains imaged from PASSCAL wide-angle seismic reflection-refraction data: Geology, v. 17, p. 119-122.

King, P.B., 1951, The tectonics of middle North America. Princeton, N.J., Princeton University Press, 203 p.

Kish, S.A., 1990, Timing of middle Paleozoic (Acadian) metamorphism in the Southern Appalachians; K-Ar studies in the Talledega Belt, Alabama: Geology, p. 18, p. 650-653.

Krol, M.A., Muller, P.D., and Idleman, B.D., 1999, Late Paleozoic deformation within the Pleasant Grove shear zone, Maryland: results from $^{40}Ar/^{39}Ar$ dating of white mica. *In* Valentino, D.W., and Gates, A.E., eds., The mid-Atlantic Piedmont: tectonic missing link of the Appalachians: Geological Society of America Special Paper 330, p.93-111.

Kunk, M.J. and Burton, W.C., 1999, $^{40}Ar/^{39}Ar$ age-spectrum data for amphibole, muscovite, biotite, and K-feldspar samples from metamorphic rocks in the Blue Ridge anticlinorium, northern Virginia [microform]: Open-File Reports, U.S. Geological Survey, 110 p.

Kunk, M.J., Horton, J.W., Jr., and Peper, J.D., 1995, Preliminary $^{40}Ar/^{39}Ar$ mineral ages from the Raleigh metamorphic belt, Carolina slate belt, and Milton belt in the South Boston, VA-NC, 30x60 minute quadrangle: Geological Society of America Abstracts with Programs, v. 27(2), p. 67.

Kutzbach, J.E., 1994, Idealized Pangean climates: sensitivity to orbital change, *In* Klein, G.D., ed., Pangea:Paleoclimate, Tectonics, and Sedimentation during Accretion, Zenith, and Breakup of a Supercontinent: Geological Society of America Special Paper 228, p. 41-55.

Kutzbach, J.E., and Gallimore, R.G., 1989, Pangean climates: Megamonsoons of the megacontinent: Journal of Geophysical Research, v. 94, p. 3341-3357.

Lapham, D.M., and Bassett, W.A., 1964, K-Ar dating of rocks and tectonic events in the Piedmont of southeastern Pennsylvania: Geological Society of America Bulletin, v. 75, p. 661-668.

Lefort, J,-P., and Van der Voo, R., 1981, A kinematic model for the collision and complete suturing between Gondwanaland and Laurussia in the Carboniferous: Journal of Geology, v. 89, p. 537-550.

Levine, J.R., and Davis, A., 1984, Optical anisotropy of coals as an indicator of tectonic deformation, Broad Top coal field, Pennsylvania: Geological Society of America Bulletin, v. 95, p. 100-108.

Lowe, D.R., 1989, Stratigraphy, sedimentology, and depositional setting of pre-orogenic rocks of the Ouachita Mountains, Arkansas and Oklahoma. *In* Hatcher, R.D., Jr., Thomas, W.A., and Viele, G.W., eds., The Appalachian-Ouachita orogen in the United States (Geology of North America, vol. F-2). Boulder, Colorado, Geological Society of America, p. 575-590.

Lukert, M.T., and Banks, P.O., 1984, Geology and age of the Robertson River pluton, *In* Bartholomew, M.J., ed., The Grenville Event in the Appalachians and related topics: Geological Society of America Special Paper 184, p. 161-166.

Lux, D.R., and Guidotti, C.V., 1985, Evidence for extensive Hercynian metamorphism in western Maine: Geology, v. 13, p. 696-700.

Mack, G.H., Thomas, W.A., and Horsey, C.A., 1983, Composition of Carboniferous sandstones and tectonic framework of southern Appalachian-Ouachita orogen: Journal of Sedimentary Petrology, v. 53, p. 931-946.

Maher, H.D., Jr., Dallmeyer, R.D., Secor, D.T., Jr., and Sacks, P.E., 1994, $^{40}Ar/^{39}Ar$ constraints on chronology of Augusta fault zone movement and late Alleghanian extension, southern Appalachian piedmont, South Carolina and Georgia: American Journal of Science, v. 294, p. 428-448.

Maher, H.D.,Jr., 1987. D3 folding in the eastern Piedmont associated with Alleghanian thrusting. *In* Secor, D.T., Jr., ed., Anatomy of the Alleghanian orogeny as seem from the Piedmont of South Carolina and Georgia, Columbia, South Carolina: Carolina Geological Society Field Trip Guidebook 1987, p. 35-48.

Maher, H.D., Jr., Dallmeyer, R.D., Secor, D.T., Jr., and Sacks, P.E., 1994, $^{40}Ar/^{39}Ar$ constraints on chronology of Augusta fault zone movement and late Alleghanian extension, southern Appalachian piedmont, South Carolina and Georgia: American Journal of Science, v. 294, p. 428-448.

Mars, J.C., and Thomas, W.A., 1999, Sequential filling of a Late Paleozoic foreland basin: Journal of Sedimentary Research, v. 69, p. 1191-1208.

Martin, W.D., and Henniger, B.R., 1969, Mather and Hockingport sandstone lentils (Pennsylvanian and Permian) of Dunkard Basin, Pennsylvania, West Virginia, and Ohio: American Association of Petroleum Geologists Bulletin, v. 53, p. 279-298.

Martin, W.D., 1998, Geology of the Dunkard Group (Upper Pennsylvanian-Lower Permian) in Ohio, West Virginia, and Pennsylvania: Bulletin of the Ohio Division of Geological Survey, 49 p.

Maystrenko, Y., Stovba, S., Stephenson, R., Bayer, U., Menyoli, E, Gajewski, D., Huebscher, C., Rabbel, W., Saintot, A., Starostenko, V., Thybo, H., Tolkunov, A., 2003, Crustal-scale pop-up structure in cratonic lithosphere: DOBRE deep seismic reflection study of the Donbas fold belt, Ukraine: Geology, v. 31, p. 733-736.

McClellan, E.A., and Miller, C.F., 2000, Ordovician age confirmed for the Hillabee Greenstone, Talladega belt, southernmost Appalachians: Geological Society of America Abstracts with Programs, v.32 (2), p. 61.

McDougall, I., and Harrison, T.M., 1999, Geochronology and Thermochronology by the ^{40}Ar/^{39}Ar Method: 2nd edition, New York, Oxford University Press, 269 p.

McLennan, S.M., Bock, B., Compston, W., Hemming, S.R., and McDaniel, D.K., 2001, Detrital zircon geochronology of Taconian and Acadian foreland sedimentary rocks in New England: Journal of Sedimentary Research, v. 71, p.305-317.

Meckel, L.D., 1967, Origin of Pottsville conglomerates (Pennsylvanian) in the central Appalachians: Geological Society of America Bulletin, v. 78, p.223-258.

Mii, H., Grossman, E.L., and Yancey, T.E., 1999, Carboniferous isotope stratigraphies of North America: Implications for Carboniferous paleoceanography and Mississippian glaciation: Geological Society of America Bulletin, v. 111, p. 960-973.

Milici, R. C., Briggs, G., Knox, L.M., Sitterly, P.D., and Statler, A.T., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Tennessee: U. S. Geological Survey Professsional Paper 1110-G, p. 1-38.

Miller, B.V., Fetter, A.H., and Stewart, K.G. (in review) Plutonism in three orogenic pulses, Eastern Blue Ridge Province, southern Appalachians: The importance of high-precision analyses in U-Pb dating of plutonic rocks: Geological Society of America Bulletin.

Miller, C.F., Hatcher R.D., Jr., Ayers, J.C., Coath, C.D., Harrison, T.M., 2000, Age and zircon inheritance of eastern Blue Ridge plutons, southwestern North Carolina and northeastern Georgia, with implications for magma history and evolution of the southern Appalachian orogen: American Journal of Science, v. 300, p.142-172.

Miller, J.D., and Kent, D.V., 1988, Regional trends in the timing of Alleghanian remagnetization in the Appalachians: Geology, v. 16, p. 588-591.

Miller, M.S., 1974, Stratigraphy and coal beds of Upper Mississippian and Lower Pennsylvanian rocks in southwestern Virginia: Charlottesville, Virginia, Virginia Division of Mineral Resources Bulletin, v. 84, 211 p.

Moecher, D.P., Cosca, M.A., and Hanson, G.N., 1997, Petrologic and $^{40}Ar/^{39}Ar$ geochronological constraints on the middle to late Paleozoic thermotectonic history of the southern Connecticut Valley zone, New England Appalachians: Geological Society of America Bulletin, v. 109, p. 164-175.

Monty, C.L.V., and Hardie, L.A., 1976, The geological significance of the freshwater blue-green algal calcareous marsh, *In* Walter, M.R., ed., Stromatolites. Elsevier, Amsterdam, p. 447-477.

Mora, C.I., Driese, S.G., and Colarusso, L.A., 1996, Middle to Late Paleozoic atmospheric CO_2 levels from soil carbonate and organic matter: Science, v. 271, p. 1105-1107.

Morton, A.C. and Hallsworth, C., 1994, Identifying provenance-specific features of detrital heavy mineral assemblages in sandstones: Sedimentary Geology, v. 90, p. 241-256.

Mueller, P.A., Heatherington, A.L.,Wooden, J.L., Shuster, R.D., Nutman, A.P., and Williams, I.S., 1994, Precambrian zircons from the Florida basement; a Gondwanan connection: Geology, v. 22, p. 119-122.

Mueller, P.A.; Heatherington, A.L.; Wooden, J.L.; Shuster, R.D.; Nutman, A.P.; and Williams, I.S. 1994. Precambrian zircons from the Florida basement: a Gondwanan connection: Geology, v. 22, p. 119-122.

Neuman, E.-R., Larsen, B.T., and Sundvoll, B., 1985, Compositional variations among gabbroic intrusions in the Oslo rift: Lithos, v. 18, p. 35-59.

Nesbitt, H.W., Young, G.M., McLennan, S.M., and Keays, R.R., 1996, Effects of chemical weathering and sorting on the petrogenesis of siliclastic sediments, with implications for provenance studies: Journal of Geology, v.104, p. 525-542.

Noel, J.R., Spariousu, D.J., and Dallmeyer, R.D., 1988, Paleomagnetism and $^{40}Ar/^{39}Ar$ ages from the Carolina slate belt, Albemarle, North Carolina; implications for terrane amalgamation with North America: Geology, v. 16, p. 64-68.

O'Connor, J.T., 1988, Carboniferous petrographic trends in the central Appalachian basin; an orogenic interpretation, *In* Schultz, A.P. (compiler) Appalachian basin symposium; program and extended abstracts: U.S. Geological Survey Open-File Report OF88-0585. p. 22-28.

O'Hara, K., and Gromet, L.P., 1985, Two distinct late Precambrian (Avalonian) terranes in southeastern New England and their late Paleozoic juxtaposition: American Journal of Science, v. 285, p. 673-709.

O'Neil, J.R., Clayton, R.N., and Mayeda, T.K., 1969, Oxygen isotope fractionation in divalent metal carbonates: Journal of Chemical Physics, v. 51, p. 5547-5558.

Olaussen, S., Larsen, B.T., and Steel, R., 1994, The Upper Carboniferous-Permian Oslo Rift; basin fill in relation to tectonic development, in Embry, A.R., Beauchamp, B., and Glass, D.J. (eds.) Pangea; global environments and resources: Canadian Society of Petroleum Geologists Memoir 17, p. 175-197.

Oliver, J., 1986, Fluids expelled tectonically from orogenic belts; their role in hydrocarbon migration and other geologic phenomena: Geology, v. 14, p. 99-102.

Osborne, T.E., 1988, Provenance and depositional environment of the Straven Conglomerate Member, Pottsville Formation (Pennsylvanian), Cahaba synclinorium, central Alabama [M.S. thesis]: Tuscaloosa, University of Alabama, 231 p.

Osborne, T.E., 1991, The depositional environment and provenance of the Straven Conglomerate Member of the Pottsville Formation, *In* Thomas, W.A. and Osborne, W.E., eds., Mississippian-Pennsylvanian tectonic history of the Cahaba synclinorium: Guidebook for the 28th Annual Field Trip of the Alabama Geological Society. p. 73-98.

Osberg, P.H., Tull, J.F., Robinson, P., Hon, R., and Butler, J.R., 1989, The Acadian orogen, *In* Hatcher, R.D., Jr.; Thomas, W.A.; and Viele, G.W., eds. The Appalachian-Ouachita orogen in the United States (Geology of North America, vol. F-2). Boulder, Colorado, Geological Society of America, p. 179-232.

Otto-Bleisner, B.L., 1993, Tropical mountains and coal formation: a climate model study of the Westphalian (306 Ma): Geophysical Research Letters, v. 20, p. 1947-1950.

Owens, B.E., and Tucker, R.D., 2003, Geochronology of the Mesoproterozoic State Farm gneiss and associated Neoproterozoic granitoids, Goochland terrane, Virginia: Geological Society of America Bulletin, v. 115, p. 972-982.

Palmer, M.R., and Edmond, J.M., 1989, The strontium isotope budget of the modern ocean: Earth and Planetary Science Letters, v. 92, p. 11-26.

Parrish, J.T., and Peterson, F., 1988, Wind directions predicted from global circulation models determined from eolian sandstones of the western United States-A comparison: Sedimentary Geology, v. 56, p. 261-282.

Parrish, J.T., 1993, Climate of the supercontinent Pangea: Journal of Geology, v. 101, p. 215-233.

Pashin, J.C., Carroll, R.E., Barnett, R.L., and Beg, M.A., 1995, Geology and coal resources of the Cahaba coal field: Geological Survey of Alabama Bulletin, v. 163, 49 p.

Petzold, D.D., 1988, Possible cyclicity of breccias within some nonmarine limestones of the Monongahela Group: Geological Society of America Abstracts with Programs, v. 20(5) p. 384.

Petzold, D.D., 1989, Nonglacial lacustrine varves from the Benwood Limestone (Pennsylvanian) of West Virginia: Southeastern Geology, v. 30, p. 193-202.

Pfefferkorn, H.W., and Thomson, M.C., 1982, Changes in dominance patterns in upper Carboniferous plant-fossil assemblages: Geology, v. 10, p. 641-644.

Potter, P.E., and Pryor, W.A., 1961, Dispersal centers of Paleozoic and later clastics of the Upper Mississippi Valley and adjacent areas: Geological Society of America Bulletin, v. 72, p. 1195-1249.

Pray, J.R., Secor, D.T., Jr., Sacks, P.E., and Maher, H.D., Jr., 1997, Rotation of fabric elements in convergent shear zones, with examples from the southern Appalachians: Journal of Structural Geology, v. 19, p. 1023-1036.

Rainbird, R.H., McNicoll, V.J., Theriault, R.J., Heaman, L.M., Abbott, J.G., Long, D.G.F., and Thorkelson, D.J., 1997, Pan-continental river system draining Grenville orogen recorded by U-Pb and Sm-Nd geochronology of Neoproterozoic quartzarenites and mudrocks, northwestern Canada: Journal of Geology, v. 105, p. 1-17.

Rankin, D.W., Drake, A.A., Jr., Glover, L., III, Goldsmith, R., Hall, L.M., Murray, D.P., Ratcliffe, N.M., Read, J.F., Secor, D.T., Jr., and Stanley, R.S., 1989, Pre-orogenic terranes. *In* Hatcher, R.D., Jr.; Thomas, W.A.; and Viele, G.W., eds., The Appalachian-Ouachita orogen in the United States (Geology of North America vol. F-2). Boulder, Colorado, Geological Society of America, p. 7-100.

Rast, N., 1989, The evolution of the Appalachian chain. *In* Bally, A.W. and Palmer, A.R., eds., An Overview: The Geology of North America A-1, Boulder, Colorado, Geological Society of America, p. 323-348.

Rast, N., and Kohles, K.M., 1986, The origin of the Ocoee Supergroup: American Journal of Science, v. 286, p. 593-616.

Rast, N., O'Brien, B.H., and Wardle, R.J., 1976, Relationships between Precambrian and lower Paleozoic rocks of the "Avalon platform" in New Brunswick, the northeast Appalachians and the British Isles: Tectonophysics, v. 30, p. 315-338.

Ratcliffe, N.M., Hames, W.E., and Stanley, R.S., 1998, Interpretation of ages of arc magmatism, metamorphism, and collisional tectonics in the Taconian orogen of western New England: American Journal of Science, v. 298, p. 791-797.

Read, C.B., and Mamay, S.H., 1964, Upper Paleozoic floral zones and floral provinces of the United States: U.S. Geological Survey Professional Paper 454-K, p. 1-35.

Reck, B.H., and Mosher, S., 1988, Timing of intrusion of the Narragansett Pier Granite relative to deformation in the southwestern Narragansett Basin, Rhode Island: Journal of Geology, v. 96, p. 677-692.

Retallack, G.J., 2001, A 300-million-year record of atmospheric carbon dioxide from fossil plant cuticles: Nature, v. 411, p. 287-290.

Rice, C.L., Hiett, J.K., Koozman, E.D., 1994, Glossary of Pennsylvanian stratigraphic names, central Appalachian basin, *In* Rice, C.L., ed., Elements of Pennsylvanian stratigraphy: Geological Society of America Special Paper 294, p. 115-155.

Richter. F.M., Rowley, D.B. and Depaolo, D.J., 1992, Sr isotope evolution of seawater: the role of tectonics: Earth and Planetary Science Letters, v. 109, p. 11-23.

Robinson, P., and Hall, L.M., 1980, Tectonic synthesis of southern New England, *In* Wones, D.R., ed., Proceedings, the Caledonides in the USA, I.G.C.P. Project 27: Caledonide orogen, 1979 meeting: Blacksburg, VA, Virginia Polytechnic Institute Memoir No. 2, p. 73-82.

Robinson, R.A.J., and Prave, A.R., 1995, Cratonal contributions to a "classic" molasse; the Carboniferous Pottsville Formation of eastern Pennsylvania revisited: Geology, v. 23, p. 369-372.

Rodgers, J., 1967, Chronology of tectonic movements in Appalachian region of eastern North America: American Journal of Science, v. 265, p. 408-427.

Rodgers, J., 1970, The tectonics of the Appalachians: Wiley-Interscience, 271p.

Rodgers, J., 1972, Latest Precambrian (post-Grenville) rocks of the Appalachian region: American Journal of Science, v. 272, p. 507-520.

Rodgers, J., 1967, Chronology of tectonic movements in Appalachian region of eastern North America: American Journal of Science, v. 265, p. 408-427.

Rodgers, J., 1987, Chains of basement uplifts within cratons marginal to orogenic belts: American Journal of Science, v. 287, p. 661-692.

Roen, J.B., 1972, Geology of the Oak Forest and part of the Blackville 7.5-minute quadrangles. I-699.

Ross, C. A., and Ross, J. R. P., 1987, Late Paleozoic sea levels and depositional sequences: Cushman Foundation for Foraminiferal Research Special Publication 24, p. 137-149.

Rowley, D.B., Raymond, A., Parrish, J.T., Lottes, A.L., Scotese, C.R., and Ziegler, A.M., 1985, Carboniferous paleogeographic, phytogeographic, and paleoclimatic reconstructions: International Journal of Coal Geology, v. 5, p. 7-42.

Sacks, P.E., Maher, H.D., Jr., Secor, D.T., Jr., and Dallmeyer, R.D., 1989, Structural transect across th Kiokee Belt and adjacent Carolina slate and Belair belts, South Carolina and Georgia: Geological Society of America Abstracts with Programs, v. 21(3), p. 56.

Sacks, P.E., Secor, D.T., Jr., Maher, H.D., Jr., Wright, J., and Dallmeyer, R.D., 1993, Early Alleghanian oblique dextral extension and magmatism along the Modoc fault zone, eastern Appalachian Piedmont, SC-GA: Geological Society of America Abstracts with Programs, v. 25 (4), p. 65.

Sacks, P.E. and Secor, D.T., Jr., 1990, Kinematics of late Paleozoic continental collision between Laurentia and Gondwana: Science, v. 250, p. 1702-1705.

Samson, S.D., Coler, D.G., and Speer, J.A., 1995, Geochemical and Nd-Sr-Pb isotopic composition of Alleghanian granites of the southern Appalachians: Origin, tectonic setting, and source characterization: Earth and Planetary Science Letters, v. 134, p. 359-376.

Samson, S.D., 2001, Timing of Alleghanian magmatism revisited: Geological Society of America Abstracts with Programs, v. 33 (2), p.7.

Samson, S.D., 1995, Is the Carolina terrane part of Avalon? *In* Hibbard, J.P., van Staal, C.R., and Cawood, P.A., eds., Current perspectives in the Appalachian-Caledonian orogen: Geological Association of Canada Special Paper 41, p. 253-264.

Samson, S.D., and Secor, D.T., Jr., 2000, New U-Pb geochronological evidence for a Silurian magmatic event in central South Carolina: Geological Society of America Abstracts with Programs, v. 32 (2), p. 71.

Samson, S.D., Secor, D.T., Jr., Hamilton, M.A., 2001, Wandering Carolina: tracking exotic terranes with detrital zircons: Geological Society of America Abstracts with Programs, v. 33 (6), p. 263.

Schedl, A., McCabe, C., Montanez, I.P., Fullagar, P.D., and Valley, J.W., 1992, Alleghanian regional diagenesis: a response to the migration of modified metamorphic fluids derived from beneath the Blue-Ridge-Piedmont thrust sheet: Journal of Geology, v. 100, p. 339-352.

Schneider, D.A., and Samson, S.D., 2000, A comparison of zircon and monazite U-Pb ages from the Rolesville Batholith, NC; lessons from misbehaving minerals: Geological Society of America Abstracts with Programs, v. 33 (2), p. 7.

Secor, D.T., Jr., Samson, S.L., Snoke, A.W., and Palmer, A.R., 1983, Confirmation of the Carolina slate belt as an exotic terrane: Science, v. 221, p. 649-651.

Secor, D.T., Jr., Snoke, A.W., and Dallmeyer, R.D., 1986, Character of the Alleghanian orogeny in the southern Appalachians: Part III. Regional tectonic relationships: Geological Society of America Bulletin, v. 97, p. 1345-1353.

Sevigny, J.H., and Hanson, G.N., 1993, Orogenic evolution of the New England Appalachians of southwestern Connecticut: Geological Society of America Bulletin, v. 105, p.1591-1605.

Shaw, H.F., and Wasserburg, G.J., 1984, Isotopic constraints on the origin of Appalachian mafic complexes: American Journal of Science, v. 284, p. 319-349.

Shelley, D., and Bossiere, G., 2000, A new model for the Hercynian orogen of Gondwanan France and Iberia: Journal of Structural Geology, v. 22, p. 757-776.

Sinha, A.K., and Zeitz, I., 1982, Geophysical and geochemical evidence for a Hercynian magmatic arc, Maryland to Georgia: Geology, v. 10, p. 593-596.

Sinha, A.K., Hanan, B.B., and Wayne, D.M., 1997, Igneous and metamorphic U-Pb zircon ages from the Baltimore mafic complex, Maryland Piedmont. *In* Sinha, A.K.; Whalen, J.B.; and Hogan, J.P., eds., The nature of magmatism in the Appalachian orogen: Geological Society of America Memoir 191, p. 275-286.

Sinha, A.K., Hund, E.A., and Hogan, J.P., 1989, Paleozoic accretionary history of the North America plate margin (central and southern Appalachians): constraints from the age, origin, and distribution of granitic rocks, *In* Hillhouse, J.W., ed., Deep structure and past kinematics of accreted terranes: IUGG 5, Geophysical Monograph, v. 50, p. 219-238.

Skehan, J.W., and Rast, N., 1990, Pre-Mesozoic evolution of Avalon terranes of southern New England, *In* Socci, A.D., Skehan, J.W., and Smith, G.W., eds., Geology of the composite Avalon terrane of southern New England: Geological Society of America Special Paper 245, p. 13-53.

Slingerland, R., and Furlong, K.P., 1989, Geodynamic and geomorphic evolution of the Permo-Triassic Appalachian Mountains, *In* Gardner, T.W. and Sevon, W.D., eds., Appalachian geomorphology: Geomorphology, v. 2, p. 23-37.

Smith, W.E., 1979, Pennsylvanian stratigraphy of Alabama, in The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States--Alabama and Mississippi. U.S. Geological Survey Professional Paper 1110-I, p. 23-36.

Smits, A.D., Stoddard, E.F., and Boltin, W.R., 1988, The Hollister mylonite zone in the northeastern North Carolina Piedmont: Geological Society of America Abstracts with Programs, v. 20 (4), p. 316.

Soreghan, M.J., Soreghan, G.S., and Hamilton, M.A., 2002, Paleowinds inferred from detrital-zircon geochronology of upper Paleozoic loessite, western equitorial Pangea: Geology, v. 30, p. 695-698.

Spear, F.S., and Harrison, T.M., 1989, Geochronologic studies in central New England I: Evidence for pre-Acadian metamorphism in eastern Vermont: Geology, v. 17, p. 181-184.

Spooner, E.T.C., 1976, The strontium isotopic composition of seawater and seawater-oceanic crust interaction: Earth and Planetary Science Letters, v. 31, p. 167-174.

Spraggins, S.A., and Dunne, W.M., 2002, Deformation history of the Roanoke recess, Appalachians, USA: Journal of Structural Geology, v. 24, p. 411-433.

Stacey, J. S., and Kramers, J. D., 1975, Approximation of terrestrial lead isotope evolution by a two stage mode: Earth and Planetary Science Letters, v. 26, p. 207-221.

Stallard, R.F., 1992, Tectonic processes, continental freeboard, and the rate-controlling step for continental denudation, *In* Butcher, S.S., Charlson, R.J., Orians, G.H. and Wolfe, G.V., eds., Global Biogeochemical Cycles: Academic Press, London, England, p. 93-121.

Stamatakos, J., Hirt, A.M., and Lowrie, W., 1996, The age and timing of folding in the central Appalachians from paleomagnetic results: Geological Society of America Bulletin, v. 108, p. 815-829.

Steltenpohl, M.G., Goldberg, S.A., Hanley, T.B., and Kunk, M.J., 1992, Alleghanian development of the Goat Rock fault zone, southernmost Appalachians: Temporal compatibility with the master decollement: Geology, v. 20, p. 845-848.

Stockey, J.R., and Sutter, J.F., 1991, $^{40}Ar/^{39}Ar$ thermochronologic evidence for late Paleozoic deformation in the Brevard zone and surrounding region near Rosman, N.C.: Geological Society of America Abstracts with Programs, v. 23 (1), p. 134.

Stovba, S.M., Maystrenko, Y.P., Stephenson, R.A., and Kusznir, N.J., 2003, The formation of the south-eastern part of the Dniepr-Donets basin: 2-D forward and reverse modeling taking into account post-rift redeposition of syn-rift salt: Sedimentary Geology, v. 156, p. 11-33.

Su, Q., Goldberg, S.A., and Fullagar, P.D., 1994, Precise U-Pb zircon ages of Neoproterozoic plutons in the southern Appalachian Blue Ridge and their implications for initial rifting of Laurentia: Precambrian Research, v. 68, p. 81-95.

Sutter, J.F., and Dallmeyer, R.D., 1982, Interpretation of $^{40}Ar/^{39}Ar$ ages from the Appalachian Grenville terrane: Abstracts with Programs Geological Society of America, v. 14 (1-2), p. 87.

Sutter, J.F., Crawford, M.L., and Crawford, W.A., 1980, $^{40}Ar/^{39}Ar$ age spectra of coexisting hornblende and biotite from the Piedmont of SE Pennsylvania; their bearing on the metamorphic and tectonic history: Geological Society of America Abstracts with Programs, v. 12(2), p. 85.

Sutter, J.F., Horton, J.W., Jr., and Kunk, M.J., 1984, Timing of Alleghanian metamorphism in the Kings Mountain belt of North Carolina and South Carolina: Geological Society of America Abstracts with Programs, v. 16(3), p. 201.

Sutter, J.F., Milton, D.J., and Kunk, M.J., 1983, $^{40}Ar/^{39}Ar$ age spectrum dating of gabbro plutons and surrounding rocks in the Charlotte Belt of North Carolina: Geological Society of America Abstracts with Programs, v. 15(2), p. 110.

Sutter, J.F., Ratcliffe, N.M., and Mukasa, S.B., 1985, $^{40}Ar/^{39}Ar$ and K-Ar data bearing on the metamorphic and tectonic history of western New England: Geological Society of America Bulletin, v. 96, p.123-136.

Tabor, N.J., and Montanez, I.P., 2002, Shifts in late Paleozoic atmospheric circulation over western equatorial Pangea: Insights from pedogenic mineral $\delta^{18}O$ compositions: Geology, v. 30, p. 1127-1130.

Talbot, M.R., 1990, A review of the palaeohydrological interpretation of carbon and oxygen isotopic ratios in primary lacustrine carbonates: Chemical Geology, v. 80, p. 261-279.

Teyssier, C., and Whitney, D.L., 2002, Gneiss domes and orogeny: Geology, v. 30, p. 1139-1142.

Thomas, W.A., 1959, Upper Mississippian stratigraphy of southwestern Virginia, southern West Virginia, and eastern Kentucky [Ph.D. thesis]: Virginia Polytechnic Institute, 322 p.

Thomas, W.A., 1966, Late Mississippian folding of a syncline in the western Appalachians, West Virginia and Virginia: Geological Society of America Bulletin, v. 77, p. 473-494.

Thomas, W.A., 1974, Converging clastic wedges in the Mississippian of Alabama, *In* Briggs, G., ed., Carboniferous of the southeastern United States: Geological Society of America Special Paper 148, p. 187-207.

Thomas, W.A., 1977, Evolution of Appalachian-Ouachita salients and recesses from reentrants and promontories in the continental margin: American Journal of Science, v. 277, p.1233-1278.

Thomas, W.A., 1988, The Black Warrior basin, *In* Sloss, L.L., ed., Sedimentary cover—North American craton, U.S.: Boulder, Colorado, Geological Society of America, The Geology of North America, v. D-2, p. 471-492.

Thomas, W.A., 1989, The Appalachian-Ouachita Orogen beneath the Gulf Coastal Plain between the outcrops in the Appalachian and Ouachita Mountains, *In* Hatcher, R.D., Jr., Thomas, W.A., and Viele, G.W., eds., The Appalachian-Ouachita Orogen in the United States, The Geology of North America, v. F-2, p. 537-553.

Thomas, W.A., 1991, The Appalachian-Ouachita rifted margin of southeastern North America: Geological Society of America Bulletin, v. 103, p. 415-431.

Thomas, W.A., and Cramer, H.R., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States-Georgia: U.S. Geological Survey Professional Paper 1110-H, p. 1-34.

Thomas, W.A., and Schenk, P.E., 1988, Late Palaeozoic sedimentation along the Appalachian orogen. *In* Harris, A.L., and Fettes, D.J., eds., The Caledonian-Appalachian orogen: Geological Society [London] Special Publication 38, p. 515-530.

Thomas, W. A., Tucker, R. D., and Astini, R. A., 2000, Rifting of the Argentine Precordillera from southern Laurentia: palinspastic restoration of basement provinces: Geological Society of America Abstracts with Programs, v. 32(7), p. 505.

Thomas, W.A., Becker, T.P., Samson, S.D., and Hamilton, M.A., 2004a, Detrital zircon evidence of a recycled orogenic foreland provenance for Alleghanian clastic wedge sandstones: Journal of Geology, v. 112, p. 23-37.

Thomas, W.A., Astini, R.A., Mueller, P.A., Gehrels, G.E., and Wooden, J.L., 2004b, Transfer of the Argentine Precordillera terrane from Laurentia: Constraints from detrital-zircon geochronology: Geology, v.32, p. 965-968.

Tollo, R.P., and Hutson, F.E., 1996, 700 Ma rift event in the Blue Ridge province of Virginia: A unique time constraint on pre-Iapetan rifting of Laurentia: Geology, v. 24, p. 59-62.

Tomascak, P.B., Krogstad, E.J., and Walker, R., 1996, U-Pb monazite geochronology of granitic rocks from Maine: Implications for late Paleozoic tectonics in the northern Appalachians: Journal of Geology, v. 104, p. 185-195.

Torsvik, T.H., Eide, E.A., Meert, J.G., Smethurst, M.A., and Walderhaug, H.J., 1998, The Oslo Rift: new palaeomagnetic and $^{40}Ar/^{39}Ar$ age constraints: Geophysical Journal International, v. 135, p. 1045-1059.

Trompette, R., 2000, Gondwana evolution: its assembly at around 600 Ma: Comptes Rendus de l'Academie des Sciences, Serie II. Sciences de la Terre et des Planetes, v. 330, p. 305-315.

Tucker, R.D., and McKerrow, W.S., 1995, Early Paleozoic chronology: a review in light of new U-Pb zircon ages from Newfoundland and Britain: Canadian Journal of Earth Sciences, v. 32, p. 368-379.

Tucker, R.D., Bradley, D.C., Ver Straeten, C.A., Harris, A.G., Ebert, J.R., and McCutcheon, S.R., 1998, New U-Pb zircon ages and the duration and division of Devonian time: Earth and Planetary Science Letters, v. 158, p.175-186.

Tucker, R.D., and Robinson, P., 1990, Age and setting of the Bronson Hill magmatic arc: a re-evaluation based on U-Pb zircon ages in southern New England: Geological Society of America Bulletin, v. 102, p.1404-1419.

Tull, J.F., 1998, Analysis of a regional middle Paleozoic unconformity along the distal southeastern Laurentian margin, southernmost Appalachians: Implications for tectonic evolution: Geological Society of America Bulletin, v. 110, p. 1149-1162.

Tullis, J., Christie, J.M., and Griggs, D.T., 1973, Microstructures and preferred orientations of experimentally deformed quartzites: Geological Society of America Bulletin, v. 84, p. 297-314.

Vai, G.B., 2003, Development of the palaeogeography of Pangaea from Late Carboniferous to Early Permian: Palaeogeography, Palaeoecology, Palaeoclimatology, v 196, p. 125-155.

Van der Voo, R., 1990, Phanerozoic paleomagnetic poles from Europe and North America and comparisons with continental reconstructions: Reviews of Geophysics, v. 28, p. 167-206.

Van der Voo, R., 1993, Paleomagnetism of the Atlantic, Tethys, and Iapetus Oceans, Cambridge University Press, 411 p.

Van der Voo, R., and Torsvik, T.H., 2001, Evidence for late Paleozoic and Mesozoic non-dipole fields provides an explanation for the Pangea reconstruction problems: Earth and Planetary Science Letters, v. 187, p. 71-81.

Valentino, D.W., and Gates, A.E., 2001, Asynchronous extensional collapse of a transpressional orogen: the Alleghanian central Appalachian Piedmont, USA: Journal of Geodynamics, v. 31, p. 145-167.

Valero Garces, B.L., Gierlowski-Kordesch, E., and Bragonier, W.A., 1997, Pennsylvanian continental cyclothem development: no evidence of direct climatic control in the Upper Freeport Formation (Allegheny Group) of Pennsylvania (northern Appalachian Basin): Sedimentary Geology, v. 109, p. 305-319.

van Houten, F.B., 1976, Late Variscan nonmarine deposits, northwestern Africa: Implications for pre-drift North Atlantic reconstructions: American Journal of Science, v. 276, p. 671-693.

Van Schmus, W.R.; and 24 others, 1993, Transcontinental Proterozoic provinces, *In* Reed, J.C., Jr.; Bickford, M.E.; Houston, R.S.; Link, P.K.; Rankin, D.W.; Sims, P.K.; and Van Schmus, W.R., eds., Precambrian: conterminous U.S. (Geology of North America vol. C-2). Boulder, Colorado, Geological Society of America, p. 171-334.

Veizer, J., Ala, D., Azmy, K., Bruckschen, P., Buhl, D., Bruhn, F., Carden, G.A.F., Diener, A., Ebneth, S., Godderis, Y., Jasper, T., Korte, C., Pawellek, F., Podlaha, O.G., and Strauss, H., 1999, $^{87}Sr/^{86}Sr$, $\delta^{13}C$ and $\delta^{18}O$ evolution of Phanerozoic seawater: Chemical Geology, v. 161, p. 59-88.

Viele, G.W., and Thomas, W.A., 1989, Tectonic synthesis of the Ouachita orogenic belt. *In* Hatcher, R.D., Jr.; Thomas, W.A.; and Viele, G.W., eds., The Appalachian-Ouachita orogen in the United States (Geology of North America vol. F-2). Boulder, Colorado, Geological Society of America, p. 695-728.

Wagner, R. and Lyons, P.C., 1997, A critical analysis of the higher Pennsylvania megafloras of the Appalachian region: Review of Palaeobotany and Palynology, v. 95, p. 255-283.

Walker, D., Simpson, E.L., and Driese, S.G., 1994, Paleogeographic influences on sandstone composition along an evolving passive margin: an example from the basal Chilhowee Group (Uppermost Proterozoic to Lower Cambrian) of the south-central Appalachians: Journal of Sedimentary Research: v. A64, p. 807-814.

Walsh, G.J., and Aleinikoff, J.N., 1999, U-Pb zircon age of metafelsite from the Pinney Hollow Formation: implications for the development of the Vermont Appalachians: American Journal of Science, v. 299, p.157-170.

Watson, E.B., and Harrison, T.M., 1983, Zircon saturation revisited; temperature and composition effects in a variety of crustal magma types: Earth and Planetary Science Letters, v. 64, p. 295-304.

Weedman, S.D., 1989, A depositional model for the upper Freeport limestone (upper Allegheny Group), Armstrong and Indiana counties, Pennsylvania, *In* Harper, J.A., ed., Geology in the Laurel Highlands of southwestern Pennsylvania: Guidebook for the Annual Field Conference of Pennsylvania Geologists, v. 54, p. 89-99.

Wegener, A., 1970, Continental drift, in Man and the Sea; Classic Accounts of Marine Exploration, American Museum of Natural History Press, Garden City, New York, p. 145-151.

Weinberg, R.F., Sial, A.N., and Mariano, G., 2004, Close spatial relationship between plutons and shear zones: Geology, v. 32, p. 377-380.

West, D.P., Jr., and Lux, D.R., 1993, Dating mylonitic deformation by the ^{40}Ar-^{39}Ar method: An example from the Norumbega fault zone, Maine: Earth and Planetary Science Letters, v. 120, p. 221-237.

West, D.P., Lux, D.R., and Hussey, A.M., II, 1993, Contrasting thermal histories across the Flying Point fault, southwestern Maine: Evidence for Mesozoic displacement: Geological Society of America Bulletin, v. 105, p. 1478-1490.

Williams, E.G., Bergenback, R.E., and Weber, J.N., 1968, Relationship between paleotopography and the thickness and geochemistry of a Pennsylvanian freshwater limestone: Journal of Sedimentary Petrology, v. 38, p. 501-509.

Williams, H., and Hatcher, R.D., Jr., 1982, Suspect terranes and accretionary history of the Appalachian Orogen: Geology, v. 10, p. 530-536.

White, A.F., Bullen, T.D., Schulz, M.S., Blum, A.E., Huntington, T.G., and Peters, N.E., 2001, Differential rates of feldspar weathering in granitic regoliths: Geochimica Cosmochimica Acta, v. 65, p. 847-869.

Whitehead, E.K., and Gromet, L.P., 1997, The Potter Hill granite gneiss, Hope Valley Terrane, SE CT; a large mid-crustal lopolith of late Paleozoic age?: Geological Society of America Abstracts with Programs, v. 29 (1), p. 89.

Whiting, B.M., and Thomas, W.A., 1994, Three-dimensional controls on subsidence of a foreland basin associated with a thrust-belt recess: Black Warrior basin, Alabama and Mississippi: Geology, v. 22, p. 727-730.

Willett, S.D., and Brandon, M.T., 2002, On steady states in mountain belts: Geology, v. 30, p. 175-178.

Willett, S.D., Fisher, D., Fuller, C., En-Chao, Y., and Chia-Yu, L., 2003, Erosion rates and orogenic-wedge kinematics in Taiwan inferred from fission-track thermochronology: Geology, v. 31, p. 945-948.

Wilson, J.T., 1966, Did the Atlantic close and then re-open?: Nature, v. 211, p. 676-681.

Wintsch, R.P., and J.N. Aleinikoff, 1987, U-Pb isotopic and geologic evidence for Late Paleozoic anatexis, deformation and accretion of the Late Proterozoic Avalon terrane, south-central Connecticut: American Journal of Science, v. 287, p. 107-126.

Wintsch, R.P., Kunk, M.J., Boyd, J.L., and Aleinikoff, J.N., 2003, P-T-t paths and differential Alleghanian loading and uplift of the Bronson Hill Terrane, south central New England: American Journal of Science, v. 303, p. 410-446.

Woodward, H.P., 1957, Chronology of Appalachian folding: American Association of Petroleum Geologists Bulletin, v. 41, p. 2312-2327.

Wortman, G.L., Samson, S.D., and Hibbard, J.P., 1998, Precise timing constraints on the kinematic development of the Hyco shear zone: Implications for the central Piedmont shear zone, southern Appalachian orogen: American Journal of Science, v. 298, p. 108-130.

Wortman, G.L., Samson, S.D., and Hibbard, J.P., 2000, Precise U-Pb zircon constraints on the earliest magmatic history of the Carolina terrane: Journal of Geology, v. 108, p. 321-338.

Wright, J.E., Sinha, A.K., and Glover, L., III, 1975, Age of zircons from the Petersburg Granite, Virginia; with comments on belts of plutons in the Piedmont: American Journal of Science, v. 275, p. 848-856.

Zartman, R.E., and Hermes, O.D., 1987, Archean inheritance in zircon from late Paleozoic granites from the Avalon Zone of southeastern New England: an African connection?: Earth and Planetary Science Letters, v. 86, p. 305-315.

Zartman, R.E., Hermes, O.D., and Pease, M.H., Jr., 1988, Zircon crystallization ages, and subsequent isotopic disturbance events in gneissic rocks of eastern Connecticut and western Rhode Island: American Journal of Science, v. 288, p. 376-402.

Ziegler, P.A., 1988, Evolution of the Arctic-North Atlantic and the Western Tethys: AAPG Memoir, v. 43, 198 p.

Ziegler, P.A., 1990, Geological atlas of western and central Europe, 2nd ed.: Avon, England, Geological Society Publishing House, Shell International Petroleum Maatschappij, 238 p.

Zoback, M.D., Zoback, M.L., Mount, V.S., Suppe, J., Eaton, J.P., Healy, J.H., Oppenheimer, D.H., Reasenberg, P.A., Jones, L.M., Raleigh, C.B., Wong, I.G., Scotti, O., and Wentworth, C.M., 1987, New evidence of the state of stress of the San Andreas fault system: Science, v. 238, p. 1105-1111.

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Publications
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Becker, T.P.; Thomas, W.A.; Samson, S.D.; and Gehrels, G.E., *accepted*, Detrital zircon evidence of Laurentian crustal dominance of the Alleghanian orogen in eastern North America: Sedimentary Geology.

O'Hara, K. and Becker, T.P., 2004, Tectonic assembly of the Brevard-Chauga belt, South Carolina: Petrological evidence from Appalachian Site Investigation Core Hole 2 (ADCOH-2): Journal of Geodynamics, v. 37, p. 565-581.

Thomas, W.A.; Becker, T.P.; Samson, S.D. and Hamilton, M.A., 2004, Detrital zircon evidence of a recycled orogenic foreland provenance for Alleghanian clastic wedge sandstones: Journal of Geology, v. 112, p. 23-37.

Abstracts:

Becker, T.P.; Thomas, W.A.; Gehrels, G.E.; and Aronson, J.L., 2004, Crustal constituency of the Alleghanian orogen as reflected by radiometric sedimentary provenance proxies: Geological Society of America, Abstracts with Programs, v. 36-6.

Becker, T.P.; Thomas, W.A.; Samson, S.D. and Gehrels, G.E., 2003, Characterization of the Late Paleozoic Alleghanian orogen by detrital zircons from the synorogenic clastic wedge: Geological Society of America, Abstracts with Programs, v. 35-6, p.389.

Becker, T.P.; Thomas, W.A.; Samson, S.D. and Hamilton, M.A., 2003, Detrital zircon evidence of proximal sources for the Alleghanian clastic wedge in eastern Tennessee: Geological Society of America, Abstracts with Programs, v. 35-1, p. 57.

Becker, T.P.; Pazzaglia, F.J.; Zeitler, P.K.; Idleman, B.D.; and Ault, A., 2001, The importance of lithology in the interpretation of apatite U-Th/He ages: Eos, AGU Transactions, v. 82, n. 47, p. 1322.

Becker, T.P., Pazzaglia, F.J., Zeitler, P.K., and Idleman, B.D., 1999, Late-stage exhumation of the central and northern Appalachians utilizing U-Th/He dating of apatite. Eos, Transactions, American Geophysical Union, v. 80, n. 46, p. 1171.

Honors and Awards

2004	Honorary Student Membership to American Association for the Advancement of Science
2002-2003	Outstanding Teaching Assistant Award Department of Geological Sciences, University of Kentucky
2002-2003	Graduate Student Research Grant Geological Society of America
2002-2003	Dissertation Enhancement Award University of Kentucky, Graduate School
2002	Pennsylvania Geological Survey Student Research Award
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Professional Positions

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